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EXOBIOLOGY IN SOLAR SYSTEM EXPLORATION

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Foreword

The advent of space exploration and modern advances in our scientific and technological capabilities are opening avenues of research never before possible. We can now search for understanding of the cosmic pathway from the formation of the elements to the condensation of the planets, from gigantic clouds of elemental gases to the synthesis of ever more complex molecules of carbon, hydrogen, oxygen, nitrogen, etc., which in turn combined to inherit the properties we have come to identify with *Life*. All of this happened in the early history of the Earth (so early that much of the record is lost) and those first self-replicating, metabolizing, mutating systems, began very quickly to interact with the planet, to modify and change the atmosphere and surface, and in turn, to evolve and adapt to the changes induced. Over the millennia, the diversity and complexity of life forms increased greatly, "intelligence" evolved and man appeared (on Earth at least) to regale himself as a crowning achievement. We know that here on the Earth DNA, proteins, etc., are essentially the same from the simplest to the most complex; bacteria to man. In the vast universe, is this planet (one of billions) actually the only one that matters, or is the evidence of the history of life to be found throughout, waiting to be deciphered? Are we an average, middle-of-the-road planet in a universe teeming with planets with and without life, similar to or vastly different from us, or are we really alone, a quirk of fate or of careful design?

Logic and some evidence suggest that we are not alone, that life must have begun many times and in many places. We can see some of the same molecules from which we ourselves are made, on dust particles in interstellar space—the ubiquity of carbon chemistry seems assured. This, however, is hardly proof. We are now able to send out robots and eventually ourselves to the other bodies in our solar system, while with our ever more powerful instruments we can dissect the cosmos far beyond our own solar system. We can capture and invade the privacy of the mysterious comets as they weave their way periodically through our community from far beyond. The evidence is at hand—we have only to collect and decipher it. It will take time and patience. The roundness of the Earth was known to the early Greeks, but was believed by many only with the perception of circling spacecraft and the ability to get far enough away from it to see it.

Mars is the first of the planets in our solar system to sustain even a rudimentary search for life. Viking sought evidence in two tiny sites and, while uncovering a wealth of data about the structure and chemistry of the atmosphere and surface, found no sign of life or even of organic molecules. Is this true of the whole planet, and if so, why? What of other planets; Jupiter or Saturn and their satellites, some nearly the size of the Earth? Have organic molecules evolved there as we believe happened here? Did life arise but not survive? What will the organic molecules of comets tell us? Are non-terrestrial processes, about which we know so little, a part of the story of life? The questions are infinite, both in scope and time, but for man to pursue his past and his future to the extent of his ability is inevitable. Meanwhile, we now have the science of "Comparative Planetology," which gives us the opportunity to study the origin and evolution of planets other than our own which will greatly improve our understanding of planetary processes so crucial to our ability to interact more equitably with our own planet.

We are no longer planet-bound. While we must come to grips with our planet and understand what we are doing to it in evolutionary terms, we are free to go beyond and apply the probe of scientific objectivity to questions heretofore assigned to the realm of philosophy. The eventual outcome will revolutionize man's basic concepts about himself and his place in the cosmos.

Richard S. Young
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Preface

During the last several years, the scope and focus of NASA's Exobiology Program has undergone an extensive re-examination as indicated in the proceedings of several workshops: "The Search for Extraterrestrial Intelligence" (NASA SP-419, 1977), "Life in the Universe" (NASA CP-2156, 1982), "The Cosmic History of the Biogenic Elements and Compounds" (NASA SP-476, 1985), "Search for the Universal Ancestors" (NASA SP-477, 1985), and "The Evolution of Complex and Higher Organisms" (NASA SP-478, 1985). Conclusions and recommendations resulting from those workshops, which addressed Earth-based studies, are now influencing the direction of the Program. Subsequently, a workshop on "Exobiology in Earth Orbit" (NASA SP-500, 1989) related many of the scientific issues raised earlier to specific mission opportunities in low-Earth orbit. Most recently, the Symposium, "Exobiology in Solar System Exploration" was held on August 24-26, 1988 in Sunnyvale, California. This Symposium provided an in-depth, comprehensive investigation of the role of Exobiology in solar system exploration. It is expected that "Exobiology in Solar System Exploration," as has "Exobiology in Earth Orbit," will provide direction for future participation of the Exobiology community in solar system exploration and alert the Planetary community to the importance of a continued, vigorous Exobiology Flight Program.

Many key questions relevant to the origin of life on Earth and to the history of the biogenic elements are best answered by studying the other eight planets and smaller bodies in the solar system. Evidence of the importance that exobiologists place on solar system exploration activities is seen in the active role they have taken in current and previous missions. The prime examples of this are the experiments that were performed on the Viking Lander Mission to Mars and the experiments and instruments that are being developed for future Mars missions. Exobiology instruments have flown on a number of planetary spacecraft and Exobiology investigators have participated in all stages of the planning and performance of the missions, as well as in the analysis and interpretation of the resulting data. Information presented at the Symposium on Exobiology in Solar System Exploration indicates that the Exobiology community's role in mission planning, conduct of flight experiments, and data interpretation should continue, and indeed, expand.

There are many solar system missions currently being planned which are of great interest to the Exobiology community. Among these are the Mars Observer, Titan-Cassini Huygen's Atmospheric Probe, Comet Nucleus Sample Return, Mars Environmental Survey, and Mars Sample Return Missions. To Exobiology's misfortune, the Comet Rendezvous Asteroid Flyby Mission fell victim to federal budget cuts and was cancelled in February 1992. However, CRAF supporters are currently examining other lower cost concepts for comet studies. In light of the upcoming opportunities, it was recommended that "Exobiology in Solar System Exploration" be held to review the current state of knowledge of the solar system with special regard to those aspects which are most relevant to Exobiology. Another objective of the Symposium was to explore and define the various roles of Exobiology in upcoming solar system exploration missions. While accomplishing these objectives, the Symposium also provided an excellent summary of the state of knowledge of planets and other solar system bodies, and of present and future solar system exploration missions. "Exobiology in Solar System Exploration" provides a sound basis for current and future participation of the Exobiology community in these missions.

Although the focus of the Symposium was primarily on Exobiology in solar system exploration missions, several ground based and Earth-orbital projects such as the Search for Extraterrestrial Intelligence, Gas Grain Simulation Facility and Cosmic Dust Collection Facility represent upcoming research opportunities planned to accommodate the goals and objectives of the Exobiology community as well.

This report contains papers for all but one of the presentations given at the Symposium. At the time these papers were written, they included the most current information available. As four years have passed since the meeting, some of the details mentioned in the papers have recently been updated by the authors. Where appropriate, we have cited references that contain additional and more current information.

We would like to express our appreciation to the Symposium speakers, along with the authors and co-authors of the papers. We'd also like to thank Douglas DeFrees for his help in planning the Symposium, taking the chair at several sessions, and reviewing the papers. Many thanks to Stanley Awramik, David Black, David Blake, Theodore Bunch, Patrick Cassen, Sherwood Chang, Donald DeVincenzi, Guy Fogleman, James Kasting, Harold P. Klein, Rocco Mancinelli, Christopher P. McKay, Verne Oberbeck, Boris Ragent, Ray Reynolds, Lynn Rothschild, Catherine Sargeant, Thomas Scattergood, Brenda Simmons, Byron Swenson, and Jose Valentin for reviewing the papers. Support by John Rummel and the NASA Life Sciences Division Exobiology Program is much appreciated. Special thanks are due to Lucille Thompson and the members of the Solar System Exploration Branch of the Space Sciences Division at Ames Research Center, without whose support this Symposium and Special Publication would not have been possible.

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Overview: Exobiology in Solar System Exploration

Glenn C. Carle and
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How, why, where, and when did life arise? What is the relationship between life and ongoing processes in the universe?

These questions are perhaps some of humankind's oldest philosophical inquiries and still resonate today. Philosophers ponder these most difficult and profound subjects from a metaphysical standpoint, while scientists have and will continue to conduct methodical studies of the physical and chemical processes of life in their pursuit to understand its origin, evolution and ubiquity.

Through the science of exobiology, we seek to understand the origin and evolution of life and life-related processes and materials throughout the universe. We know a great deal about contemporary living systems and the way they function, but as we ask about earlier and

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earlier life forms, there are fewer known facts. Much has been learned about exobiology, but considerably more knowledge is left to be gained before the subject is well understood.

Understanding the relationship between the origin and evolution of life and the origin and evolution of the solar system is one of the main objectives of exobiology. Clearly, life is intimately connected with its environment. With respect to the origin and subsequent evolution of life on Earth, the environment played a critical role. When life first arose on Earth, the biogenic elements (carbon, hydrogen, nitrogen, oxygen, phosphorus, sulfur) and traces of many other elements must have been available in appropriate amounts. Temperatures must have been moderate enough for water to exist as a liquid, but not so hot as to destroy organic molecules. The environment must have been conducive to prebiotic organic chemistry (i.e., the evolution of simple chemicals to complex molecules necessary for development of living systems); otherwise, the prebiotic precursors to life could not have formed. Much about the interplay between environment and prebiotic process remains to be understood. Some of the questions exobiologists seek to answer are: What were the sources and complexities of the prebiotic molecules? What role did the environment play in mediating or directing the chemical evolution that fostered Earth's first replicating system? Is Earth the only place in the solar system (or in the universe, for that matter) where life exists? Could life have originated and survived elsewhere?

While pursuing the answers to these weighty questions, exobiologists extrapolate from the single example they have access to: life on Earth. They study contemporary biology in the laboratory by way of the conventional disciplines such as microbiology, biochemistry, and biophysics. They also study the Earth's geological and fossil record by way of the disciplines of geology and paleontology. Earth itself complicates this work. The ancient rock record, which may have once held the key to chemical evolution and the origin of life, has apparently been destroyed by ongoing geological processes. The oldest fossil evidence of life are stromatolites, which date back to about 3.5 billion years ago. Although only slightly younger than Earth's oldest rocks, these life forms are very similar to the contemporary microbial mat communities often found in shallow waters. The fossil stromatolites were, as the microbial mats are today, complex communities comprising a number of different types of microorganisms. While the fossil stromatolites are very ancient, they were highly evolved ecosystems and certainly do not represent Earth's earliest life forms.

Given that evidence of the earliest epochs has not yet been discovered and may not be available on Earth, exobiologists have turned outward to the rest of the solar system to understand the conditions that fostered life. Furthermore, in other bodies of the solar system, they may find additional environments where life now exists or where it once existed. Even though the Viking spacecraft found no evidence of life existing on Mars, if early Mars was like the early Earth, life may have begun on Mars, also. Fortunately, substantial areas on Mars have had little or no geological recycling since its earliest period. Sediments containing fossils, or other biomarkers, have not been subducted, ground up, or remelted as they have been on Earth. Therefore, the possibility exists for a preserved geological record of the origin of life on Mars. In fact, Mars may hold the only existing geological evidence of the origin of life in the solar system. A mission to Mars with exobiology-specific experiments and instruments may find this evidence.

Exobiologists also seek knowledge of the earliest environment on Earth by investigating the role that comets may have played in delivering to Earth biogenic materials necessary for chemical evolution and the origin of life. Scientists ask, What role did cometary impacts play in the origin of life? What was the composition of Earth's atmosphere during stages of chemical evolution and the origin of life? According to one theory, comets introduced essential organics into the atmosphere spurring development of biopolymers. Recent data from encounters with Halley's Comet support these claims, showing that comets contain substantial amounts of water and organic compounds. The exact nature of these organics is as yet unknown and their contributions to chemical evolution are also unknown. Missions to other comets may provide information needed to understand Earth's earliest chemical and physical processes.

Continuing Inquiries

In 1958, the Congress of the United States enacted the National Space Act. This act validated the importance of exploring space and established, for the first time, an organization that could provide direct access to the solar system. This organization was the National Aeronautics and Space Administration (NASA). In 1984, the Subcommittee on Space Science and Applications reviewed the Space Act and concluded that

> *A healthy space science program is essential to continued strength and vitality of the space program...(and that) NASA should initiate enhancements in the space and earth sciences including pursuit of planetary exploration through a renewed commitment to exploration of the solar system, and expansion of human knowledge of the Earth and phenomena in the atmosphere and space.*

NASA has carried out many important steps in this pursuit. Two notable accomplishments are the Apollo Missions with the concurrent search for life on the Moon and the Viking Missions which searched for life on Mars. While neither investigation found evidence for extraterrestrial life, each gathered a wealth of information about these and other bodies in the solar system. This information is still being interpreted. Today, many new opportunities lie on the horizon. In the United States and the Commonwealth of Independent States (formerly the Soviet Union), as well as in Europe and Japan, missions are being planned to comets, the outer planets and their satellites, and Mars. It is appropriate, therefore, to reassess the current status of knowledge about the bodies of the solar system and identify the role currently planned missions can play in expanding our understanding of exobiology.

In August 1988, the NASA Ames Research Center held a three-day symposium in Sunnyvale, California, to discuss the subject of exobiology in the context of exploration of the solar system. Leading authorities in exobiology presented invited papers and assisted in setting future goals. The goals they set were to:

- Review relevant knowledge learned from planetary exploration programs;

- Detail some of the information that is yet to be obtained;

- Describe future missions and how exobiologists, as well as other scientists, can participate; and

- Recommend specific ways exobiology questions can be addressed on future exploration missions.

These goals are in agreement with those of the Solar System Exploration Committee (SSEC) of the NASA Advisory Council. Formed in 1980 to respond to the planetary exploration strategies set forth by the Space Science Board of the National Academy of Sciences' Committee on Planetary and Lunar Exploration (COMPLEX), the SSEC's main function is to review the entire planetary program. The committee formulated a long-term plan (within a constrained budget) that would ensure a vital, exciting, and scientifically valuable effort through the turn of the century. The SSEC's goals include:

- Determining the origin, evolution, and present state of the solar system;

- Understanding Earth through comparative planetology studies; and

- Revealing the relationship between the chemical and physical evolution of the solar system and the appearance of life.

The SSEC's goals are consistent with the over-arching goal of NASA's Exobiology Program, which provides the critical framework and support for basic research. The research is divided into the following four elements:

1) Cosmic evolution of the biogenic compounds;

2) Prebiotic evolution;

3) Origin and early evolution of life; and

4) Evolution of advanced life.

NASA's program is designed to trace evolutionary pathways of the universe, including synthesis of biogenic elements, evolution of planetary systems, origin of life, and evolution of intelligent life.

To attain these goals, relevant studies of the planets and other bodies of our solar system are necessary. The SSEC's strategy for exploration is to begin with a global view of the solar system bodies, and only later move to detailed observations and measurements in selected regions. Although much information is gained from first generation reconnaissance missions, *in situ* atmospheric and surface measurements and the analysis of returned samples provide the specific details needed for interpretation and understanding. Such missions also assist us in modeling the processes that occur on planets and smaller solar system bodies.

The Inner Planets

The inner planets and the Moon have been explored more intensely than the outer planets or small bodies due to their proximity and similarity to the Earth. Missions have included flyby and orbiter missions, soft-landings of spacecraft on Mars, probes into the Venusian atmosphere, and manned missions to the Moon with samples returned.

The Moon

The U.S. Lunar Program began in 1964 with Ranger VII's photography-based mission, and culminated in 1972 with the Apollo project. Apollo landed men on the Moon, returned lunar samples from six sites, and permitted a quantum leap in our understanding of spaceflight and lunar evolution. Although life apparently did not evolve on the Moon, studies help us to better understand Earth and conditions and events that occurred more than 3.5 billion years ago.

Compared with Earth, the Moon is depleted of hydrogen, carbon, nitrogen and their simple compounds. Small amounts of water and other volatiles, as well as minuscule amounts of carbon found in fine-grained material, may be vestiges of impacting bodies. No clear evidence of complex organic compounds has been found, but traces of methane and carbide-like species have. This suggests a synthesis from interactions between the solar and the lunar surfaces. In fact, solar wind provides a major portion of the Moon's hydrogen, carbon, and nitrogen. Investigators found that the parts per billion (ppb) amounts of amino acids in lunar soils were not indigenous, but rather produced during analyses from trace amounts of chemical precursors of amino acids. Apparently, solar and cosmic irradiation, meteorite bombardment, and volatile loss have severely constrained the Moon's chemical behavior and the evolution of biochemically significant elements. Questions still to be answered include, How old are the lunar impact craters? Did impacts occur in any particular pattern, or were they random? What are the compositions of meteorites on the lunar surface?

If we plan to use lunar resources to further planetary exploration, we first need to understand the Moon in very close detail. During the proposed Lunar Orbiter Mission, data will be collected to compile a global lunar map. From this data we should be able to deduce the composition of the lunar surface, the presence of condensed water, and other volatiles. It has been suggested that a lunar base would be the ultimate location to study the origin and history of the Moon, which may better help us understand the early history of the Earth. From this base investigations using the radio-quiet far side could search for evidence for extraterrestrial intelligence and conduct other astronomical investigations. Additionally, scientific investigations could take advantage of the Moon's high vacuum and extreme changes in temperature.

Venus

Venus is too hostile today to harbor life; temperatures are very high as clouds cover 100% of the surface. If life did exist, it did so far in the past when conditions were more Earth-like. The United States began studying the planet Venus in 1962 with the launch of the first successful interplanetary probe, Mariner II. In addition to this and Pioneer Venus, various Soviet Venera missions have provided data concerning the properties of the surface of Venus. Although controversial, data collected during these missions suggest the possibility that ancient lakes may have once spread across the surface of Venus. Models also imply that Venus may have possessed a sizable water inventory, as well as a cooler climate, in its early history.

These putative lakes (or oceans) could have remained for several hundred million years, until they boiled away as temperatures rose because of a runaway greenhouse effect. There is also a suggestion that comets, thick with ice, could have replenished those waters for nearly a billion years. The possible presence of lightning may have been an important energy source for the production of organic chemical compounds. Early environmental conditions on Venus may have been favorable for the development of life, but if the impact flux was as high as it was on Earth, early life on Venus may have been repeatedly wiped out. The study of Venus may give us insight into the physical limitations of the habitability of an Earth-like planet.

Presently, data collected during the Magellan Mission are being used to create a global map of Venus. Magellan entered orbit around Venus in April 1990. The probe's mapping radar has already mapped over 95% of the planet and has revealed that about 85% of the planet is covered by volcanic rocks, mostly lava flows that form the great plains. Detailed studies of the surface features and their characteristics will allow us to address questions on the age, origin, and history of Venus.

Mars

In 1976 and 1977, after several Mariner flybys and after many remote images had been taken of the Martian surface, two Viking Landers touched down on Mars. Because the search for life on Mars was a major driver for this Mission, three microbiological experiments were performed. However, no unequivocal evidence was found for metabolic activity in any of the surface samples analyzed, no organic matter was detected, and no images were returned suggesting signs of life. These data led us to believe the possibility of extant life on Mars was quite slim, unless it existed in hidden, protected niches.

However, even if there is no extant life on Mars, it is still reasonable to hypothesize that life may have once existed there. Geologic and climatologic studies suggest that the early Martian environment may have been similar in many respects to that of the early Earth. Both planets show histories of liquid surface water, warmer temperatures, and cometary and meteoritic impacts, as well as relatively thick carbon dioxide and nitrogen atmospheres and volcanic activity. Therefore, the elements needed for chemical evolution may have been present on Mars, even though subsequent planetary events did not favor the continuing evolution of life.

Regardless of whether life originated on Mars, the planet may contain within its sedimentary record the early histories of both Mars and Earth. Because of Earth's geologic processes, unaltered rocks that date back to 3.5-4 billion years ago are rare. On Mars, however, over half the planet (the southern hemisphere) dates back to the late bombardment event of about 3.8 billion years ago. Studying Mars, therefore, not only helps us understand events that occurred while life on Earth was originating, it also allows us to investigate the possibility of an extinct Martian biota.

The study of Mars provides us with clues to better discriminate between conflicting theories of solar system formation, revealing to us new clues about how terrestrial planets with atmospheres formed and evolved and allowing us to more fully understand planet Earth. New missions to Mars will study Martian atmospheric and surface materials and processes, and will search below the planet's surface for buried clues of past, and perhaps even present, biological activity.

One planned mission is the Mars Observer Mission (MO), scheduled for launch in September 1992. The Mission will continue more detailed studies of the Red Planet by sending a spacecraft platform to orbit Mars. Remote sensing instruments will observe and map the entire Martian surface and atmosphere for at least one Martian year, collecting data on surface materials, volatiles and dust; topography and the gravitational and magnetic fields; and the structure and circulation of the atmosphere. MO data will constitute a baseline for future Mars missions. MO's most important contribution to exobiology will be improving our understanding of climate change and its consequences on any past biota and biological processes.

The Mars Environmental Survey (MESUR) Network Mission is planned for several launch opportunities beginning in 1999. The Mission objective is to establish a global network of stations on the surface of Mars to concurrently collect and return scientific data over a minimum of one Martian year. The full network will consist of 16 stations providing pole-to-pole coverage of Mars. Each station will be identical, weigh less than 150 kilograms, and house instrumentation to characterize the planet's meteorology, internal seismic activity, and local surface properties. Using data obtained during the Mission, exobiologists will gain insight into the geological and atmospheric conditions in the past and present, and Mars' ability to sustain a potential past biota. MESUR is viewed by the scientific community as the logical, evolutionary step after Viking and before sample return/human exploration missions.

The Mars Rover Sample Return (MRSR) mission is currently proposed for the late 1990s. Recommended by the SSEC and the NASA's Office of Exploration Task Group, chaired by Sally Ride, the MRSR mission will: gather data to better understand the origin and evolution of Mars and the conditions present on very early Mars; search for evidence of former life; and improve our knowledge of the Martian environment in preparation for human exploration. The Rover will sample different materials at different locations and will perform a range of *in situ* measurements. Because the MRSR mission will only return 5-10 kilograms of material, selected samples must represent the planet's full variety. Once returned to Earth, the samples can be analyzed with the benefit of tools and techniques, such as sample dating and trace element analyses, that are too complicated to be performed remotely. Data taken during the MRSR mission will help answer questions about how the surface of Mars evolved, what the structure of the interior is, whether or not life once existed on the planet, what the planet's present thermal and dynamic states are, and what role water played in the evolution of the Martian surface.

The Outer Planets

The outer planets and small bodies have been explored less intensely than the inner planets and the Moon. A recent accomplishment in space exploration is the successful reconnaissance of Voyager 2 with all the giant planets during its grand tour of the solar system.

The giant planets include Jupiter, Saturn, Uranus, Neptune, and Pluto. They formed in the same flattened disk of gas and dust, the solar nebula, as did the terrestrial planets, but are strikingly different. Thought to have retained much of the original solar nebular material, the giant planets are composed of three basic materials: gas (mostly hydrogen and helium), ice (containing mixtures of water, carbon and nitrogen-containing materials), and rock (mixtures of silicon, magnesium, iron, oxygen, and other heavy elements).

The giant planets are bigger and more massive than the terrestrial planets, and have numerous rings and moons surrounding them. The giant planets themselves exhibit a wide variation in basic properties. Atmospheres, for example, extend ten thousand to several tens of thousands of kilometers into the planets' interiors, thereby representing a significant fraction of a planet's entire mass. These atmospheres are highly reducing environments, where methane is converted into more complex molecules.

Some exobiologists believe the chemical steps leading to life involved the production of complex carbon-containing molecules from simple molecules in a highly reducing environment. If this is so, then atmospheres of the giant planets represent natural laboratories for observing and understanding initial chemistry that leads to production of organic molecules. Energy sources driving the atmospheric chemistry include solar ultraviolet radiation, lightning, and high-energy charged particles that are precipitated into auroral zones. Lightning may be important in producing certain disequilibrium species such as hydrogen cyanide and carbon monoxide. If carbon and other biologically relevant materials came to Earth from the outer solar system, the giant planets may have played a fundamental role in transferring this material to the inner solar system by virtue of their gravitational interactions with small bodies.

Jupiter

Our understanding of life as a planetary phenomenon is based upon Earth-like planets. However, there are environments in the solar system where liquid water, commonly believed to be a prerequisite for biological activity, may exist in a distinctly non-Earth-like environment. Such an environment may occur on Jupiter's moon, Europa. There, surface water ice, organic molecules and biogenic elements (available as precursors for the synthesis of biocompounds), are thought to exist along with a continuous supply of energy (either through the dissipation of tidal energy as heat or through hydrothermal activity or solar radiation).

Europa formed from the Jovian nebula, rich in water and other biogenic elements; therefore, early Europa may have been favorable for the capture and *in situ* formation of different organic compounds, including biochemical monomers and polymers. The low-temperature environment may have also facilitated the preservation and further interaction of labile biochemical polymers. Whether these processes proceeded to a more advanced stage is not known; no experimental work has been done at low temperatures on the protocellular stages of prebiological evolution. Some exobiologists believe that if any Europan biota exists, it would have to be anaerobic and prokaryotic.

The Galileo Mission will take a close look at Jupiter and Europa. It was launched from the Space Shuttle in October 1989 and follows a complex looping orbit to arrive at Jupiter in late 1995. Once in the Jovian vicinity, it will spend many years examining the Jovian moons and Jupiter's atmosphere.

Laboratory research indicates that Jupiter's copper and gold colors are most likely due to an atmosphere rich in biogenic elements, such as phosphorus, methane, ammonia, and hydrogen gas. Discovery of large storms on Jupiter during the Voyager project set the stage for a natural laboratory for the study of chemical evolution. The Galileo Mission will allow us the opportunity to take advantage of this natural laboratory once again.

The Galileo Mission will also help elucidate questions of exobiological interest: Is there abundant organic material present on Europa? Do the fractures on the smooth, icy surface indicate the presence of liquid water beneath the surface? Are there internal sources of energy that would promote life? Can prebiotic chemistry occur at such cold temperatures? The monitoring of Europa for vapor and frost clouds that could result from fracturing events may reveal whether liquid water lies beneath its icy surface. Radar sounding could also detect liquid water if it is present. High resolution imaging of Europa's surface could elucidate processes involved in the formation of its linear features, as well as in variations of its coloration and albedo.

Future missions using landing spacecraft to penetrate the surface and perform *in situ* surface and subsurface measurements in Europa will be required to find out what actually lies beneath the surface.

Saturn

Saturn's rings are composed of particles ranging in size from tiny pebbles to large boulders and include a sprinkling of fine dust. The material is mostly icy, and contains impurities of unknown origin. The Saturnian satellites have dark coatings presumed to be carbon-rich organic material. The source of these coatings is unknown, but further analysis may help to elucidate the relationship among icy satellites, comets, and primitive carbonaceous meteorites. By continuing studies of the outer planets and by comparing their structure and behavior, we gain insight into the formation and evolution of the solar system and can learn what chemistry might have taken place here on Earth during the first mysterious years.

Titan is Saturn's largest satellite. Titan is an icy, rocky object that possesses a dense nitrogen-methane atmosphere. Although its size and atmospheric density might stimulate comparisons with terrestrial planets, at such a great distance from the Sun, Titan is much colder and richer in ices. The surface temperature is about 94K, only a few degrees above the melting point of methane. It has been postulated that lakes or oceans of liquid methane or other hydrocarbons might dominate the surface, and that dense methane clouds exist in the lower atmosphere below the visible orange haze.

Although the conditions on present day Titan are quite different from those that are thought to have existed on early Earth, Titan provides a natural setting, on a planetary scale, for the study of processes involved with chemical evolution. The low temperatures found on Titan preclude the existence of significant amounts of water vapor; however, abiotic chemical evolution, leading to complex hydrocarbons and nitriles, can still take place. Also, Titan's hazes and ubiquitous colored clouds indicate the presence of aerosols. These aerosols can interact with the atmospheric gases and with each other, thereby influencing the path of atmospheric evolution. Chemical analyses of Titan's atmosphere should provide more information about the primordial nebula from which Titan was formed.

Information about material presently impacting Titan may also be obtained. Because hydrogen cyanide has been detected in Titan's atmosphere, syntheses of as yet undetected organics may have occurred in the atmosphere. Clearly, synthesis of some abiotically useful compounds should have taken place, albeit anhydrously. This is likely to have been the case, as evidenced by compounds already detected. Therefore, careful inventories of the various organic compounds in the atmosphere of Titan are essential to expand exobiology's knowledge of prebiotic evolution of the precursor molecules necessary for the origin of life. Many questions remain unanswered: What kind of prebiotic chemistry is taking place in Titan's lower reaches? Do these processes provide clues as to how precursor molecules led to life on Earth?

The Cassini Mission is a vital component of an exciting exploration program. While much of what we know about Saturn's and Titan's atmospheres has come from Voyager flyby data, the Cassini Mission to Saturn and Titan will address fundamental questions about the formation and evolution of the solar system, prebiological organic chemistry, plasma physics, atmospheric dynamics, and virtually every other aspect of space science. Data retrieved from Saturn may provide insight into the dynamics and nature of Saturn's ring particles, Saturn's atmospheric composition and the nature of its magnetosphere. Additionally, Saturn's moons will be explored and the surface composition and geologic history of these moons will be determined. Together with the Galileo Mission to Jupiter, these missions represent a systematic and comprehensive approach to the study of the outer planets and will provide tremendous insight into the formation and structure of the solar system.

The Cassini Mission will consist of two parts. The first will be the delivery of a probe into the atmosphere of Titan. The probe will sample and analyze chemical and physical characteristics of the atmosphere and haze particles in detail. The probe will be followed by a four year orbital tour of all parts of the Saturnian system, including numerous flybys of Titan and the other main "icy" satellites (Mimas, Enceladus, Tethys, Dione, Rhea, Hyperion, and Iapetus).

NASA will build the orbiter as the second spacecraft in its Mariner Mark II series; these are modular spacecraft designed for exploration of the outer planets and small bodies. The European Space Agency (ESA) will provide the Titan Huygens Probe, which will be launched with the orbiter. The Huygens Probe will begin sampling the atmosphere and transmitting data at an altitude of about 170 km. The descent from this level to the surface will last approximately 2.5 hours. Because of the thick haze layer, the surface of Titan has never been observed; its precise nature will remain unclear until the Huygens Probe is able to resolve it.

Comets, Asteroids, and Other Objects

Comets are known to contain a relatively large amount of organic molecules. They are vehicles potentially capable of delivering to the planets products of interstellar processes, such as organic compounds produced in space by nonequilibrium reactions (catalysis and radiation processing of condensed volatiles). Despite their apparent diversity, comets represent a population of essentially homogeneous objects that show strong analogy with carbonaceous chondrites.

Comets eventually decay into gas and dust, becoming a source of interplanetary particles containing carbon. Comets may also be a major source of planetary atmospheres. It is believed that Earth may have received most of its ocean and organic molecules from heavy comet bombardment as the solar system formed. These simple and complex organic molecules may still be preserved in comets today. Despite invaluable data gathered from the Comet Halley flyby, many questions remain: What is the complete inventory of organic materials present in comets? How has this material been synthesized into complex, biologically interesting molecules? Is there any evidence that substantial amounts of material from comets have been delivered to the planets?

The concept of orbital diffusion throws a new light on exobiology. It suggests the existence of a cosmic mechanism, working everywhere, that can transfer prebiotic compounds to ubiquitous rocky planets so as to provide a proper environment to initiate life. However, not enough is known about cometary chemistry for a full understanding. The study of Comet Halley has brought a wealth of new information, but the fast flyby velocity destroyed significant molecules that were thus not characterized. An active, short period comet that has entered the inner solar system only very recently is the best target for exobiology studies. Such a comet should still contain a substantial supply of volatiles, and a high level of activity should assure a sufficient amount of gas and dust reaching the spacecraft for studies.

Exobiology is also interested in the continuing study of asteroids and meteorites. Most asteroids, particularly those toward the outer parts of the asteroid belt, are objects of very low-surface reflectivity. Similarly, the surfaces of cometary nuclei appear to be very black. Some planetary satellites are also black on a global scale or contain black matter mixed with surface ices. Macromolecular carbon, found in meteorites and interplanetary dust particles (IDPs), is a plausible low-albedo constituent of these dark (asteroid, comet, and satellite) surfaces.

The presence of organic matter in cometary dust has been established. How much organic material do asteroids contain? To what extent, and how, has the organic material in asteroids and meteorites been synthesized into complex, biologically interesting molecules? The source of the organic matter in the meteorites and asteroids is unclear, but isotopic composition (deuterium enrichment) indicates formation in low-temperature interstellar molecular clouds. Organic synthesis, in the form of photochemical smog, is currently in progress in some planetary atmospheres (Jupiter, Saturn, Uranus, Titan) and on planetary and satellite surfaces (Pluto, Triton), producing yellow and brown-colored organic complexes from the methane that is detected spectroscopically. Alteration of these organics to more complex solids by the action of ultraviolet and cosmic ray flux probably occurs. Telescopic searches for the CH spectral signature strongly suggest its presence on dark, hydrous asteroids, but confirmation and further exploration are needed.

Questions still remain about the origin of meteoritic organic matter, particularly concerning the role played by presolar, interstellar processes and factors controlling the extent of chemical prebiotic evolution on different solar system bodies. Using data obtained during the October 1991 Galileo encounter of Gaspra, and the planned, fast flyby of Ida in August 1993, we will gain insight into the nature of asteroids and the composition and source of their low-albedo component(s).

Dust is a ubiquitous component of our galaxy and solar system. Interstellar dust is the predominant form of condensable elements in the galaxy that are not in stars. Grains form in gas outflows from stars and they are processed, destroyed and reformed in the interstellar medium and molecular clouds. Interplanetary dust is debris recently liberated from comets and asteroids within the solar system and is relatively rich in volatiles, a source of abiotic organic molecules. The interstellar medium offers a rich variety of environments in which carbon chemistry can occur. A wide variety of chemical processes provides for the production of rich assemblages of biogenically interesting compounds.

A strong link exists between interplanetary and interstellar dust. Prior to the formation of the solar nebula, most atoms heavier than helium were contained in interstellar grains. Some grains were incorporated into comets and asteroids. Both interstellar and interplanetary dust seeded early Earth with elements and compounds. Collection and analysis of extraterrestrial dust through the use of dust-collection facilities on Earth-orbiting vehicles will provide information about the sources of biogenically significant elements and compounds that accumulated in distant regions of the solar nebula and that were later accreted onto planets.

Also contributing to the understanding of interstellar and interplanetary dust is the Long Duration Exposure Facility (LDEF). LDEF is a large, reusable, unmanned, free-flying spacecraft accommodating technology, science, and applications experiments for long-term exposure to the space environment. The LDEF was placed in Earth orbit in April 1984 by Challenger, and retrieved from orbit in January 1990 by Columbia. Several experiments on board the LDEF studied the chemistry and morphological characteristics of micrometeoroids and interplanetary dust. Data obtained during those experiments should provide an understanding of the biogenic significance of such extraterrestrial materials.

Due to federal budget cuts, the Comet Rendezvous Asteroid Flyby (CRAF) Mission was canceled in February 1992. If resurrected, the CRAF Mission will perform measurements aimed at a deeper understanding of chemical evolution. It will address fundamental questions about the formation and evolution of the solar system, prebiological organic chemistry, plasma physics, and atmospheric dynamics. This Mission will also help provide evidence as to whether substantial amounts of cometary material have been delivered to the planets, particularly Earth.

CRAF will study a typical short-period comet, Kopff, over more than half its orbit, following aphelion near the orbit of Jupiter through perihelion and beyond. In 2000, CRAF will reach its destination with Comet Kopff and be inserted in an orbit around the Sun that precisely matches that of the comet. CRAF will map the entire surface of Kopff with imaging and spectroscopic instrumentation. It will identify different ices and minerals on the comet's surface, and will use accurate radio tracking to determine the mass of the comet. The spacecraft and comet will then travel together through one or more complete orbits. Detailed analyses will be made of the gas and dust flowing from the comet's nucleus. A year after the spacecraft's arrival, a penetrator carrying a set of instruments will be placed directly into the surface of the nucleus to measure elemental composition, temperature, and surface strength and to determine the physical state and composition of the icy organic mix below any crust on the surface.

En route to comet Kopff, CRAF will fly by and observe a primitive, carbonaceous-type, main-belt asteroid, 449 Hamburga, and make remote sensing measurements. Data obtained will address questions concerning elemental and isotopic abundances, chemical composition, ice phases, and the mineralogic forms and microscopic physical structures of these primitive solar system materials which are of great interest to exobiologists.

The CRAF Mission will provide an opportunity for significant advances in the knowledge of primitive (yet complex) material in short-period comets and the surface of an asteroid. The Mission will obtain new information concerning the early environment of the solar system and the extent of chemical evolution in such environments.

Comparison of data with results of analyses of interplanetary dust and meteorites will elucidate relations between these different solar system bodies, and should give a clearer picture of how primitive bodies, comets, asteroids, and interplanetary debris may have contributed to the origin of life on Earth.

The data anticipated from CRAF will yield much insight on comets, but ultimately a comet sample return mission, such as the proposed Rosetta Sample Return Mission, will be needed to provide the range and detail of analyses needed to definitively characterize a comet nucleus.

Summary

Many fundamental questions remain about the origin of life, chemical evolution, and the formation and composition of our solar system. A number of spacecraft missions have been proposed that will certainly provide very important clues and answers to some of these outstanding questions. For exobiology, the highest priorities for future missions should include a thorough investigation of a comet and a future cometary nucleus sample return mission. In addition, it is recommended that a comprehensive study of Mars be pursued and that a sample return mission to Mars, with intelligently chosen samples from carefully selected sites, be a high priority.

Future exploration objectives should be to determine the present nature of the solar system, its planets and primitive bodies to understand how the solar system and its objects formed, evolved, and (in at least one case) produced an environment that could give rise to and sustain life. Another objective should be that of comparative planetology: to better understand the Earth by determining the general processes that govern all planetary development and by understanding why terrestrial planets differ so much from each other.

National goals for space exploration were outlined in the Space Exploration Initiative (SEI). First announced in 1989 and established by the Bush Administration in 1990, SEI provided the basic framework for future exploration: back to the Moon and then a piloted mission to Mars. The 90-Day Study, 1989, contained explicit suggestions for the successful exploration of the Moon and Mars. More recently, the Synthesis Report, 1991, was the first serious look at how to fulfill the President's stated objectives in space. The Report contains the findings of a year-long survey of the U.S. aerospace engineering community and the general public, and features four basic options for returning humans to the Moon and proceeding toward Mars. These options include:

- A trip to Mars in 2014 by six astronauts, with testing of related technology on and around the Moon beginning in 2005;

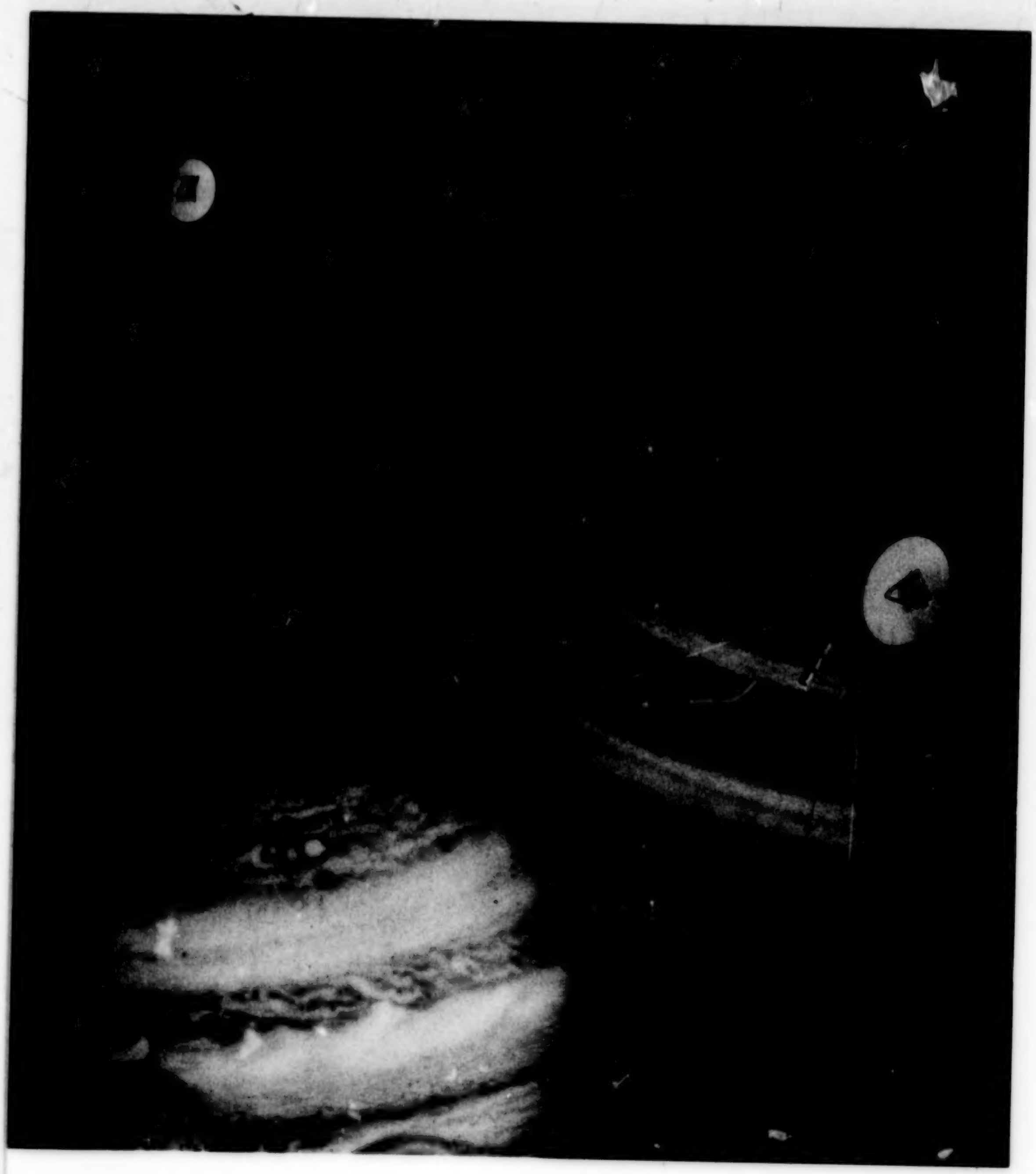
- Scientific exploration of the Moon and Mars, with the first piloted lunar landing in 2003 and extensive robotic activity on both bodies;

- A permanent Moon base followed by human excursions to Mars;

- A Moon-oriented approach toward returning solar energy and other natural resources to Earth, with a self-sufficient Mars base initiated in 2016.

To achieve these objectives and advance exploration of the solar system at the healthy pace recommended by the SSEC, we must strengthen our ground-based planetary science research effort, put energy into major technological developments, build up our research and data analysis capabilities, and increase our potential for telescopic observations and remote analysis capabilities. Additionally, we need to develop and foster relationships with other nations to further international cooperation.

With a national commitment to human presence in the solar system endorsed, and reports such as the Synthesis Report and the Ride Commission Report "Leadership and America's Future in Space" recommending Mission to Planet Earth, future lunar bases, expanded planetary exploration, a series of Mars Sample Return missions, and the human exploration of Mars, we are preparing for a future of national leadership and activity beyond Earth orbit and into the solar system.



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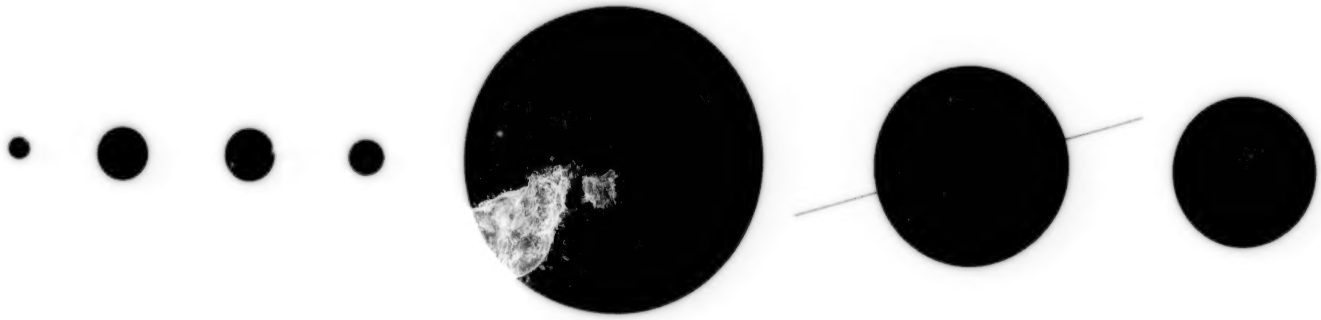

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The Solar System: Importance of Research to the Biological Sciences

H. P. Klein

Even as the two Viking Landers touched down to begin their experiments on the Martian surface in 1976, many of those who were skeptical about finding life on Mars were proclaiming that the biologists associated with space research would soon be "out of business"; that the search for extraterrestrial life would prove to be negative (at least within our solar system); and that, therefore, there would be no further interest in pursuing the subject of exobiology. Indeed, as the data from the Vikings accumulated and were analyzed, the prospects for extant life on Mars became remote, as did the likelihood of studying extraterrestrial organisms.

It was almost 30 years ago that Joshua Lederberg coined the term "exobiology" to describe the then-new scientific enterprise (more or less loosely defined as the study of extraterrestrial life). While Lederberg made it abundantly clear that the actual search for extraterrestrial life was

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part of the larger question of how life arises from inanimate matter, to many, exobiology came to be synonymous with the *search* for extant life on other planets. Such was, and still is, the view of many individuals who do not understand the biological context within which the search for life on Mars was carried out.

What needs emphasis is that the Viking Mission was, for biologists, an important test of our ideas on how life arises from primordial nonbiological materials. In retrospect, we were probably naive in our assumptions about the ease with which replicating organic systems are formed and evolve on a planetary body. The Viking results have emphasized how much more we need to know about the history of chemical and biological evolution.

Exobiology today involves scientists from a wide variety of disciplines—from astronomy to molecular biology—all contributing to the common goal of understanding how matter in the universe evolves toward replicating systems—i.e., toward life. In essence, we are trying to understand the *course of evolution*—going backward in time to the origins, in stars, of the biogenic elements necessary for life; through the subsequent processing of these elements in the interstellar medium and on planetesimals and planets, to yield organic compounds; and, ultimately, to the appearance of replicating organic molecules. In extensions of this evolutionary theme, exobiology also encompasses inquiry into the earliest stages of *biological evolution* and how the evolution of biological systems is affected by the physical evolution of a planetary body.

It is important to realize that the goal of understanding the course of chemical and biological evolution will not only require intensive study of the Earth and a more complete history of terrestrial organisms, *but will also require critical data that exist elsewhere in the solar system.*

The Role of Space Missions

During the past two decades, we have seen important advances in the study of chemical evolution which provide ample laboratory experimental verification that very simple compounds can be transformed into organic matter of reasonable complexity. However, we are on tenuous grounds when we try to pin down these evolutionary processes in time and space.

In this regard, solar system missions have been important in helping us to understand the early environment of the Earth and the subsequent evolution of organisms. But there are still wide gaps in our knowledge of the processes that led to the origin of life and conflicting models of the environments within which biology, once initiated, first began to evolve. Many of these uncertainties can be substantially constrained by continued exploratory missions to objects in the solar system, several of which have surfaces and atmospheres that still retain information—lacking on the Earth—about the early history of the solar system.

History of the Biogenic Elements and Their Compounds

In tracing the history of organic chemical evolution prior to the formation of the solar system, one way in which space missions can contribute to this understanding is by the study of interplanetary dust particles. What is their composition? What is their relationship to interstellar and circumstellar grains which are believed to aggregate and ultimately condense to form new planets, and which are also known to contain organic compounds? It is important to understand any evolutionary relationships among these types of grains. Evidence has already been obtained suggesting that grains originating from pre-solar system sources can become incorporated into solar system planetesimals such as meteorites.

Characterization of the molecular complexity in these grains will require observations in the far infrared, visible, and ultraviolet regions of the electromagnetic spectrum. In addition, critical spectral regions will require that observations be made using space-based instruments with high angular and spectral resolutions.

Another important area that can benefit from space missions is the actual collection of dust grains, taking advantage of devices on board a space station. In principle, it should be possible to separately collect interstellar and interplanetary particles for subsequent study. Furthermore, the possibility of collecting these in a nondestructive manner, to preserve volatiles and organic compounds, is an important objective for such investigations.

As we continue to trace the fate of grains containing the biogenic elements and organic compounds, it is important to learn how the incorporation of such grains and associated gases into the solar nebula may have affected preexisting organic compounds. Processes such as pyrolysis and evaporation could have caused their loss or destruction, while processes such as turbulence and radiation may have promoted the synthesis of organics. Almost certainly, further chemical modifications took place as the nebular dust grains accreted into larger objects, such as comets, asteroids, and planets. What were the chemical processes that may have affected the biogenic materials during this phase of the formation of bodies in the solar system? To obtain insight into these processes, it would be highly desirable to understand the formation of grain aggregates. Studying some aspects of this process in the absence of gravity (for example, in a laboratory on a space station) can be a useful technique in this area.

Vital clues to the history of the biogenic elements also exist today in primitive solar system objects, including meteorites, comets, and asteroids. What are the detailed compositions of these bodies? What are the distributions of organic compounds in them? What interrelationships exist among the various classes of objects? What presolar phases can be identified in them? To answer these questions adequately, direct access is needed to these objects. This, in turn, will require a variety of *in situ* and sample return solar system missions. Among these, highest priority should be given to investigations of comets, particularly those comets likely to have preserved intact nebular gases and grains containing the biogenic elements and their compounds.

Early Planetary Environments and the Origin of Life

As we contemplate the processes on the primitive Earth leading to the origin of life, one major question revolves around the nature of the primordial atmosphere. How reducing were conditions on the Earth during that period? Was the atmosphere dominated by hydrogen, thus facilitating the abiotic synthesis of organic precursors necessary for life? For this, the giant outer planets—with their hydrogen-rich atmospheres—can provide a wealth of information on the nature of the chemistry that is possible under highly reducing natural conditions. For this reason, these planets represent prominent targets for exobiological study. What is the distribution of organic matter in these planets? More significantly, what is the origin of these compounds? To elucidate these questions, detailed *in situ* measurements will be required. For example, knowledge of the distribution with altitude of organic matter, carbon monoxide, and phosphine in the atmospheres of Jupiter and Saturn—coupled with computer and laboratory simulation experiments—can lead to a better understanding of the chemistry that is taking place on these bodies.

The Saturnian satellite, Titan, is a special case for study in the outer solar system. Here too, extensive organic chemistry must be occurring, as evidenced by the large number of organics that have been identified in its atmosphere. What level of complexity is reached by organics on Titan? What is the chemistry that accounts for these? Without a doubt, atmospheric entry missions, such as the "Titan-Cassini" Mission, will be required in order to answer many of the questions about the organic components on Titan.

In addition to the dark surface revealed by the recent comet Halley encounter, dark coatings are also found on the surfaces of Phoebe, Iapetus, some of the outer-planet satellites, on some asteroids, and in the rings of Uranus. On all of these bodies there is thus the suggestion of carbonaceous matter. What is the relationship of these to each other and to the dark, carbonaceous, surface matter of meteorites rich in organic compounds, such as the Murchison meteorite? What is their mode of formation? For the resolution of these questions, it will be necessary to obtain considerable new and sophisticated spectroscopic data concerning their composition. Some may come from ground-based observatories, but a more complete understanding will require spaceborne instruments and, ultimately, *in situ* and sample return missions.

One of the large unknowns surrounding the environment within which the first organisms arose has to do with the possible contribution by impacting bodies to the Earth's inventory of volatiles and organic compounds. It is essential that we obtain better insight into this issue. When and how were the volatile materials necessary for life added to the Earth's surface? To assess the role of cometary and asteroid impacts in the early history of the Earth, many problems need to be solved. We need to put better bounds on the input of cometary and asteroidal volatiles and organics, we need to model the size distribution of such impacting bodies on the primitive atmosphere and the effects of such impacts on the synthesis and destruction of organics, and we need to integrate this information with the most reasonable models of the primitive atmosphere with regard to its redox characteristics.

These issues will require theoretical and experimental studies of the physics of cometary and meteoritic impacts in order to place limits on the survival rates of organic compounds, as well as thermochemical calculations of the composition of atmospheres produced by large impacts. High-sensitivity visible and infrared searches for comets in the Oort cloud would be highly desirable to determine the large end of the size distribution of objects that may have impacted the early Earth.

At present it is impossible to estimate the contribution of organic matter to the prebiotic Earth by impacting bodies. Equally difficult is the assessment of the nature of any organics that might have survived this process. Were the important precursor molecules for life formed subsequently? If so, how were the precursors of biological macromolecules synthesized on the prebiotic Earth? What mechanisms were involved? What were the first stages of molecular replication?

Of necessity, work in this area requires modeling the composition and history of the early terrestrial atmosphere and evaluating the early terrestrial environment with regard to other parameters, such as temperature, atmospheric pressure, solar radiation, etc.

For meaningful laboratory simulations of possible prebiotic processes for the synthesis of precursor molecules, the simulations must be carried out under conditions that are consistent with the most realistic models of the Earth at that time. *In turn, these models will be strongly dependent upon information gathered from solar system missions.*

Early Biological Evolution

When did the first stable biota appear on this planet? It is not at all clear whether the initial form of life may have been wiped out by some catastrophic event, only to arise again later—perhaps even several times—before more clement conditions were sustained on the Earth.

What environmental constraints must be placed on the early Earth for determining the timing of the origin of life and for the *maintenance* of the first organisms? Answers to these questions again will entail studies of the effects of impacting bodies on the post-accretionary Earth. Significant insights are also to be obtained from modeling studies to determine the biological effects of "rock vapor" and of "tsunamis" created by large-body impacts. Corollary analyses of lunar and Martian cratering data will help to provide limits on the size and frequency of such early large-body impacts. Additional critical geophysical and geochemical data will also be necessary, requiring analysis not only of ancient terrestrial rock samples, but also of rock samples from Mars.

Evolution of Cellular and Multicellular Life

Another area of evolutionary biology that is strongly dependent upon models of the terrestrial environment is one that deals with the evolution of the major biological groupings. It is now abundantly clear that biological evolution is closely linked to the physical evolution of this planet and that to understand biological evolution one needs to integrate the biological record with the physical record of the planet. What was the nature and what were the times of origin of the major groups of prokaryotic and eukaryotic organisms? What was the sequence of evolution of these organisms? Can we correlate the phylogenetic relationships among organisms with the geological record? Understanding how physical environmental factors may have driven the *early* biosphere will require detailed analysis of well-preserved sedimentary basins, particularly those of Archean and Proterozoic ages, for integrating the biological and geological records. Better constraints are needed on the temperature history of the Earth, on the history of the atmosphere, and on the chemistry of the oceans, as well as on the history of solar radiation. Once more, studies in this field will depend upon valid models of the evolving Earth, and these will be derived, at least in part, from comparative planetology and from studies of the Sun.

Can we find, in the geological record, convincing evidence that cosmic events have influenced the course of biological evolution? To address this issue, studies are needed that will be directed toward the recognition of extraterrestrial (e.g., asteroidal, cometary) "signatures" in terrestrial sedimentary successions, as well as research to evaluate temporal patterns in the fossil record.

Importance of Mars

Of all the objects for study beyond the Earth, Mars assumes a position of prime importance for exobiological study. Based on knowledge derived from ground-based observations and from previous space missions, a vast amount of information has accumulated concerning the gross environmental characteristics of Mars. These indicate an environment that is relatively Earth-like when compared to other objects in the solar system. The information upon which this assessment is made is far from complete. Furthermore, considerably less is known about the early history of Mars. Nevertheless, it is reasonable to assume that Mars was accreted at about the same time, out of essentially similar materials, as was the Earth, and that the subsequent cosmic influences on Mars (e.g., solar radiation and cometary and asteroidal impacts) closely paralleled those felt by Earth.

Given these general considerations, it can be asked whether organic chemical evolution proceeded on that planet in its early history along lines similar to those presumed to have taken place on the Earth. What isotopic, molecular, morphological, and environmental evidence for chemical evolution and the origin of life can be found on Mars?

Missions to Mars, coupled with theoretical studies, will be needed to reconstruct the history of liquid water and its interactions with surface materials, and to model the early geochemical cycles of carbon, nitrogen, sulphur, and iron. Missions are also needed, along with laboratory and theoretical model studies, to elucidate photochemical and weathering processes on Mars and the nature of inorganic carbon, nitrogen, and iron-bearing phases in the Martian surface materials. The nature of the oxidant revealed by the Viking experiments also needs to be determined, and further searches for organic compounds need to be made if we are to understand the course of chemical evolution on Mars.

It is also important to note, in connection with issues relating to the evolution of cellular and multicellular forms of life, that one aspect of the exploration of Mars is that it affords us, in principle, an opportunity to determine whether any fundamental inferences about the earliest stages of biological evolution on the Earth can be extended to Mars. That is, can we generalize our understanding about early cellular evolution on Earth by comparative studies of Mars?

In the near term, investigations into this area of research will require both *in situ* and sample return missions to Mars. Ultimately, field studies on that planet will almost certainly be necessary to maximize the opportunities for evaluating the early environment and the potential for cellular evolution on Mars.

Life Outside the Solar System

Finally, this discussion cannot be complete without some reference to the subjects of chemical evolution and the origin of life on planets *outside* our solar system.

Clearly, all of our theories ascribing the origin of life to the gradual evolution of replicating molecules as part of the process of *planetary* evolution coupled with the growing conviction that other planetary systems can be spawned in the universe, lead to the presumptive conclusion that living systems will have arisen elsewhere. To the exobiologists this possibility is considered to be quite plausible, since it represents a logical extension of our basic contentions. For this reason research directed toward finding and characterizing other planetary systems is of special significance.

Conclusion

In the foregoing brief review, an attempt has been made to describe the scope of scientific areas that comprise the current field of exobiology in the United States. From investigations of astrophysical phenomena that deal with the birth of stars and planetary systems to questions of molecular biology involving phylogenetic relationships among organisms, from attempts to simulate the synthesis of biological precursor molecules in the chemistry laboratory to making measurements of the organic constituents of Titan's atmosphere, these researches all converge toward a common objective—answering a question that humans have been asking since the dawn of their existence: "Where did we come from?"

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The Moon: Biogenic Elements

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E. K. Gibson, Jr. and S. Chang

Among the highest research priorities for life sciences in space is the search for extraterrestrial life and for the fundamental understanding of life's origins. This search, coupled with the investigation of the origin and evolution of the solar system, constitutes the essential element of a quest for a "cosmic perspective" for humankind; a quest that may illuminate our long evolutionary path from elemental origins in the primal fireball, to formation of galaxies, through interstellar dust clouds, solar systems, and planets (inorganic chemical evolution), to the first stirrings of life in the Earth's primitive oceans (organic chemical evolution), and the beginnings of biological evolution. Humanity's explorations of the cosmos are essential steps in the development of this new perspective.

The origin and evolution of life is tied to the conditions of primitive planetary environments. Throughout the history of the space program it has been assumed that there is a connection between evolutions of life and planets. Until humans went to the Moon in 1969, the only extraterrestrial material available for study was the meteorites which fall randomly on the Earth. The early search for direct evidence of life and/or organic chemical

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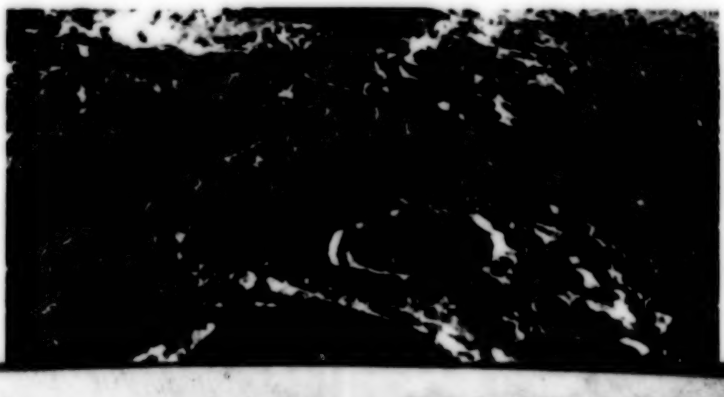

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evolution in extraterrestrial bodies was limited to Earth-bound laboratory studies of meteorites; no evidence of extraterrestrial life was detected, and terrestrial contamination confounded any adequate assessment of the real significance of organic compounds isolated from meteorites. Since then, the organic chemical analyses of meteorites and the study of the interstellar medium through radio telescopes have revealed the presence of organic compounds of unambiguous extraterrestrial origin, thereby providing clear evidence that organic chemistry occurred or does occur elsewhere in the solar system and cosmos.

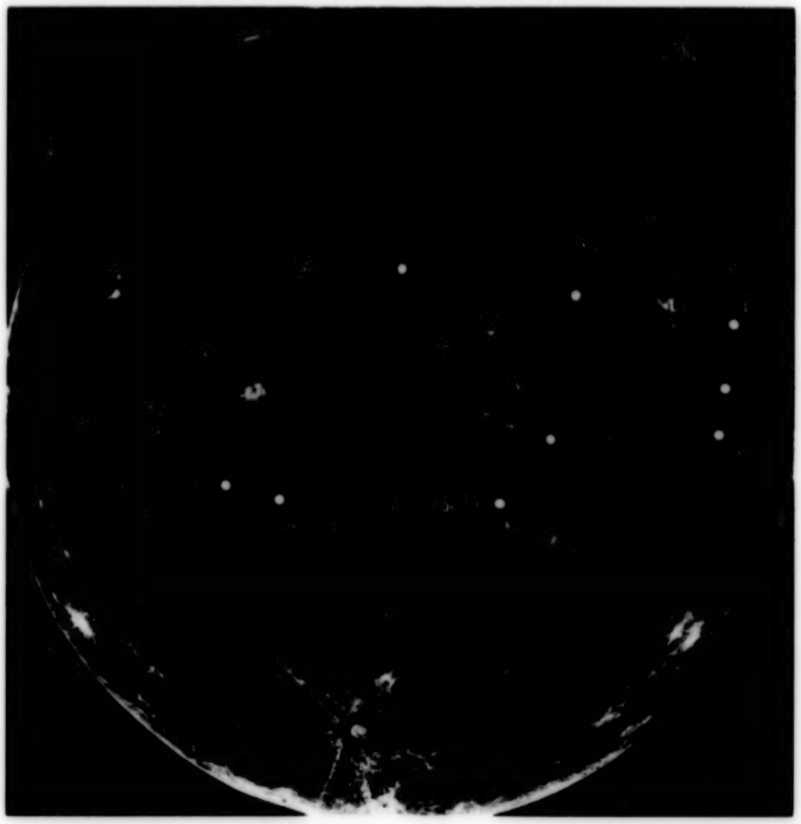

Figure 2-1. Map of the Moon showing the Apollo and Luna landing sites where samples were collected and subsequently returned to Earth for detailed study.

The specific objectives of the organic chemical exploration of the Moon involve the search for molecules of possible biological or prebiological origin. Detailed knowledge of the amount, distribution, and exact structure of organic compounds present on the Moon is extremely important to our understanding of the origin and history of the Moon, and to its relationship to the history of the Earth and solar system. Specifically, such knowledge is essential for determining whether life on the Moon exists, ever did exist, or could develop. In the absence of life or organic matter, it is still essential to determine the abundance, distribution, and origin of the biogenic elements (e.g., H, C, O, N, S, P) in order to understand how the planetary environment may have influenced the course of chemical evolution.

During the early 1960s various investigators suggested that the lunar surface might contain elevated carbon concentrations and even abundant organic matter. The first direct measurement of a possible carbon abundance for the lunar surface was obtained by the Surveyor missions which soft-landed spacecraft on the lunar surface during 1966-68. On board three of the Surveyor probes were alpha back-scattering instruments which measured between 0.2 and 0.9 ± 0.2% carbon. However, the analytical error in the measurements was so large that a true carbon abundance for the lunar surface remained unknown until the return of samples by the manned Apollo missions.

From 1969 through 1976 samples were returned from nine sites on the Moon (fig. 2-1). The six Apollo sites offered samples collected from a wide variety of localities by the twelve astronauts. The Soviet Luna program landed three automatic sample collecting robots which collected soil samples with a drill located at the base of the landers, and subsequently returned the samples to Earth. The six Apollo lunar missions returned 2,196 samples, weighing a total of 381.69 kg (841 pounds). The three Luna missions from the Moon returned three core samples with a combined weight of 300 gms (0.6 pounds). The lunar samples were examined in pristine laboratories prior to release to investigators throughout the world for detailed study (fig. 2-2).

Within the Lunar Receiving Laboratory (LRL) and other laboratories, over 3,000 tests for viable organisms were carried out with different nutrient media, at various temperatures, and in different atmospheric mixtures of oxygen, carbon dioxide, and nitrogen. No viable life forms were found, not even terrestrial contaminants. Micropaleontological studies of the lunar soils gave no evidence of structures resembling terrestrial microfossils that could be attributed to past biota. In retrospect the noted astronomer, C. Huygens, was correct when he wrote in 1757 that, "the Moon has no air or atmosphere surrounding it as we have, [and I] cannot imagine how any plants or animals whose whole nourishment comes from fluid bodies can thrive in a dry, waterless, parched soil."

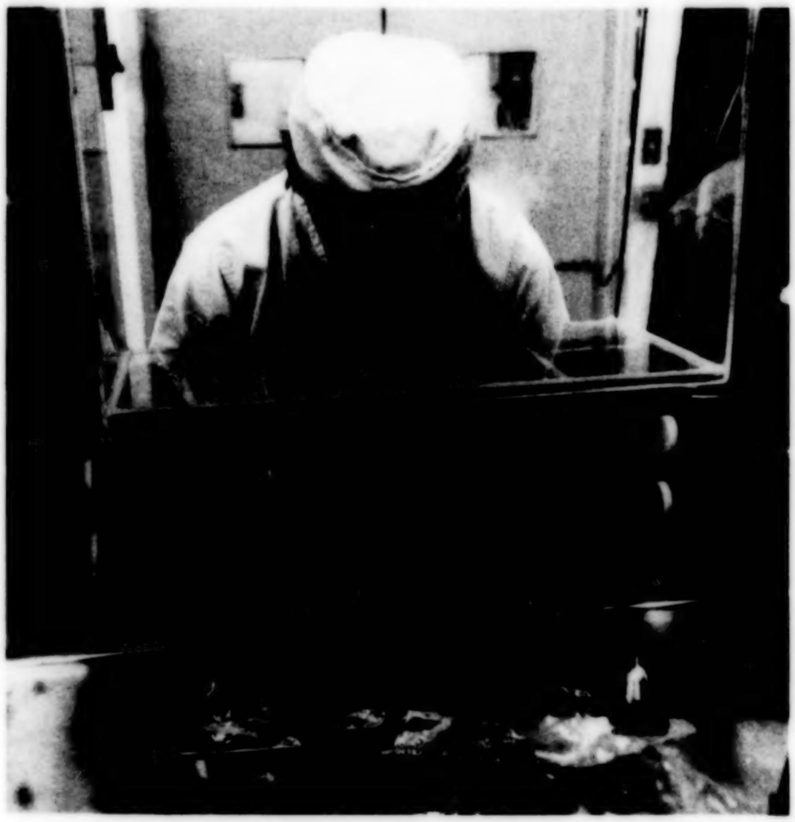

Figure 2-2. Lunar Receiving Laboratory scientist examining a large Apollo 14 sample. Extreme care was used in the processing of samples so that minimal contamination was introduced. Stainless steel, aluminum, and teflon were the only materials which came in contact with the samples in a sample processing cabinet filled with purified dry nitrogen. Despite such precautions, contamination by terrestrial carbon and other elements still occurred. When extreme precautions were taken to prevent contamination of the lunar samples, clean samples were delivered to the principal investigators. For example, quantities of carbon as small as 1 microgram were measured in lunar rocks and this work assisted in identifying a spallation carbon component present in lunar materials.

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The detailed analysis for organic compounds involved the separation of minute quantities of organic constituents from the inorganic lunar sample matrix, separation of the resulting mixtures by various types of chromatography into classes of compounds such as acidic, basic, and neutral, polar and nonpolar, followed by separation of each group into individual components. Instrumental techniques sensitive to the parts per billion (ppb) level were employed to determine the detailed structures and amounts of the resulting compounds.

Extensive studies carried out on returned lunar samples to find complex organic compounds which might be precursors for biochemistry failed to obtain any unequivocal evidence of such species at the ppb level. The parts per million (ppm) amounts of high molecular weight hydrocarbons extracted from lunar soils in several instances were apparently not of lunar origin, but could be accounted for as contaminants introduced during laboratory processing of the samples. Methane and the simple C_2 and C_3 hydrocarbons have been found in trace amounts in lunar soils. The methane, at least, is believed to be of lunar origin (see below). The ppb amounts of amino acids found in lunar soils were not indigenous to the Moon, but were produced during the analyses from trace amounts of chemical precursors of amino acids. Follow-on tests suggested that these precursors were contaminants present as rocket exhaust products within the lunar soils. These results led to the conclusion that if synthesis of complex organic compounds ever occurred on the Moon, very little evidence remains. Nonetheless, C, H, and N are found in lunar samples, and intensive study of distributions and origins of these elements has shown them to be of extralunar origin.

Lunar soils have carbon abundances ranging from 4 to 280 ppm by weight (table 2-1). The average total carbon for 178 lunar soils from 6 Apollo missions is 115 ppm, while the mean is 78 ppm C for 57 breccias (formed by impact heating and compaction of lunar soils), and 10 crystalline rocks contained on the average less than 2 ppm. The work on crystalline igneous rocks using stepwise oxidation and isotopic measurements (see below) showed that the Moon as a whole is strongly depleted in carbon, as it is in all other volatiles including some metallic elements. Apparently, abundance data for carbon in lunar rocks reported by others, primarily reflected the presence of terrestrial contaminants.

Table 2-1: Lunar Biogenic Element Abundances
(in parts per million)

	Soils	Breccias	Rocks
Hydrogen			
Range	3.2-60	12-48	1.2-3.8
Mean	40	68	2.1
Carbon			
Range	4-280	30-250	0.1-3
Mean	115	78	1.5
Nitrogen			
Range	7-164	16-181	1-6
Mean	82	68	3
Sulfur			
Range	290-1400	20-2250	20-2800
Mean	850	756	1950 basalts 300 anorthosites

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The carbon concentration in any soil is dependent upon its exposure history while on the lunar surface. Immature lunar soils (short exposure time) generally have carbon contents below 40 ppm. The unusual and highly immature orange soil from Apollo 17, which has been shown to be a volcanic glass, contains a wide range of carbon (4 to 100 ppm), but the variability probably reflects different amounts of terrestrial contaminants rather than differences in adsorbed carbon from lunar volcanism. Soils characterized as mature by various indices of exposure on the lunar surface such as solar wind argon content, exposure age, and agglutinate content (see below) contain the highest carbon contents.

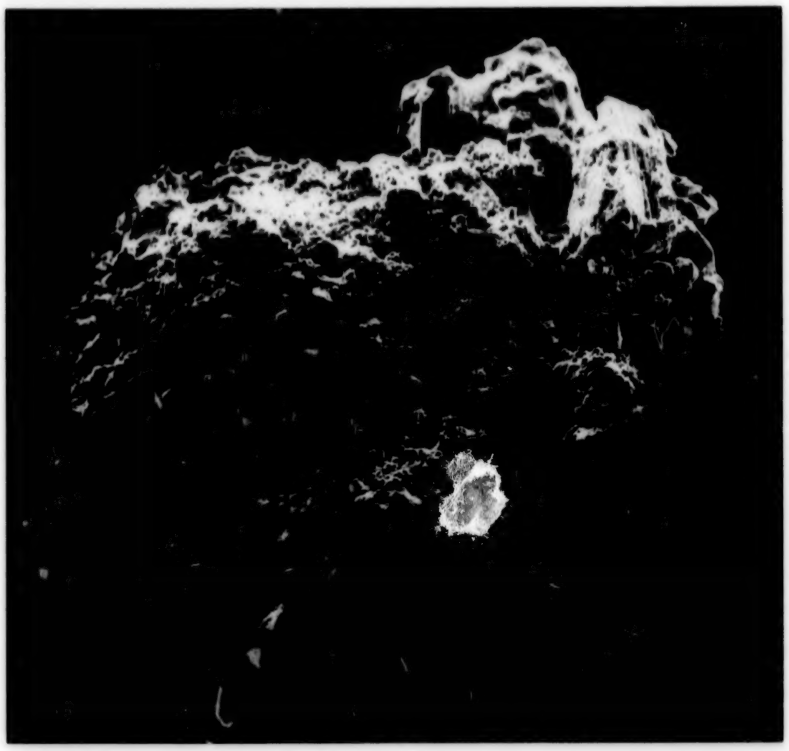


Figure 2-3. Photomicrograph of a lunar agglutinate particle. The agglutinate is a constructional particle produced during micrometeorite and/or meteorite impact on the lunar surface. The particle consists of lithic, mineral, and glassy debris bonded together by an inhomogeneous glass. The particle is approximately 1 mm in size.

Selected lunar soils have been separated into their individual components and studied for their carbon contents. Within a lunar soil, agglutinates (glassy components consisting of fine-grained, comminuted rock, mineral, and glass fragments bonded together with glass) produced during meteorite or micrometeorite impacts (fig. 2-3) contain the highest carbon concentrations. The breccia fragments contain intermediate amounts, while the individual mineral grains contain the least carbon. In soil components separated by grain size, the highest concentrations of carbon occur in the smallest grains, and the concentrations are inversely correlated with grain size indicating that the accumulation of carbon is related to the surface area of the grains.

Carbon at the lunar surface is contained primarily in the fine-grained material of the regolith where layers of dusty ejecta have accreted over time as a result of meteoritic impacts (fig. 2-4). The present distribution of carbon and other volatile elements in the regolith is a consequence of a dynamic equilibrium between addition and loss processes associated with solar radiation and meteorite bombardment (fig. 2-5). The lunar carbon cycle is governed by these regolith processes, in contrast to the carbon cycle on Earth which is mediated by volcanism, sedimentation, and biological activity. Since crystalline rocks contain only a few ppm carbon at most, and a significant atmosphere cannot have been retained on the Moon, indigenous lunar sources such as volcanism and the attrition of the crystalline rocks cannot have contributed much to the content of the soils. Indeed, the preponderance of data so far (see below) supports the view that carbon is contributed primarily from sources outside the Moon such as the solar wind, meteorites, and comets. Among these sources, the evidence favors the solar wind on the basis of isotopic composition and because hypervelocity (≥11 km/s) impacts of meteorites and comets with the Moon are expected to vaporize the projectile and some of the target with subsequent loss of the gases to space. Possibly the interaction of the solar wind with this pulse of gases may result in charge exchange and reimplantation of a small fraction of the vaporized material into the surface of grains in the regolith. Some estimates place the contribution of meteoritic or cometary carbon to a mature lunar soil at 5 to 10 ppm.

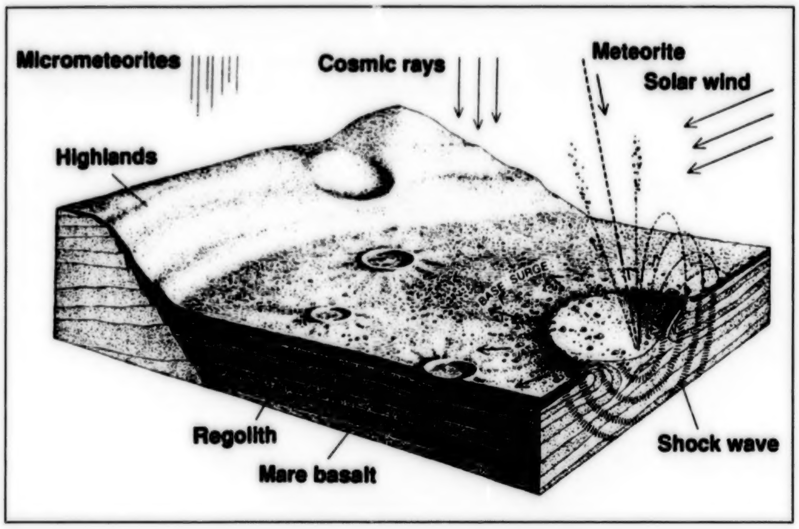


Figure 2-4. Processes of erosion on the surface of the Moon are extremely slow compared with the processes on Earth. Bombardment by micrometeorites is the main cause, removing approximately 1 mm of the surface of the rocks in a million years. A large meteorite strikes the lunar surface very rarely, excavating bedrock and ejecting it over thousands of square kilometers, sometimes as long rays of material radiating from the resulting crater. Much of the meteorite itself is vaporized on impact, and larger fragments of the debris create secondary craters. Such an event at a mare site pulverizes and churns the rubble and dust that form the regolith. Accompanying base surges of hot clouds of dust, gas, and shock waves might compact the dust into breccias. Cosmic rays continuously bombard the surface. During lunar day ions from the solar wind and unshielded solar radiation impinge on the lunar surface. (After Eglinton et al., 1972.)

The solar wind is a stream of charged particles (mainly ionized hydrogen with the other elements in their solar abundances) moving outward from the Sun with velocities in the range 300-500 km/sec. For every 7,500 hydrogen nuclei in the solar wind, there is one carbon nucleus. The ionized particles interact

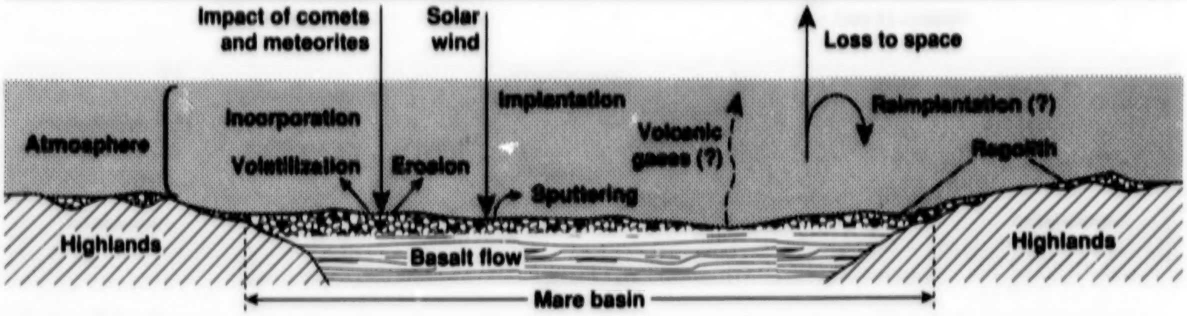


Figure 2-5. Biogenic elements at the lunar surface are contained primarily in the regolith of the maria and in the lunar highlands where layers of dust have slowly accreted. It is not expected that layers of coarse ejecta contain much carbon or biogenic elements, and the crystalline igneous and highlands anorthositic rocks contain only a few parts per million of the biogenic elements. The present distribution of carbon and other biogenic elements in the regolith is a consequence of a dynamic equilibrium. Hydrogen, carbon, and nitrogen are contributed from sources outside the Moon such as the solar wind, meteorites, and comets, and from lunar sources such as volcanism and the attrition (wearing down) of the crystalline rocks. Possibly some of the carbon released into the tenuous lunar atmosphere produced by sputtering or volatilization of impacting projectiles is reimplanted. The biogenic elements hydrogen, carbon and possibly nitrogen are lost to the atmosphere by diffusion, erosion and volatilization. They are constantly lost from the Moon to space, sometimes very rapidly, as during the impact of a meteorite. The volatile element abundances of the lunar soils measured now represent the end result of more than three billion years of these processes. The dust of the lunar maria may be regarded as both the reservoir and the reaction site for the biogenic elements at the surface of the Moon. (After Eglinton et al., 1972.)

directly with the lunar surface and are imbedded into the outer portions (i.e., 500 angstroms thickness) of lunar soil components and rocks which are exposed to the Sun. Solar wind irradiation of planetary surfaces occurs only if the planetary body does not have a magnetic field of sufficient strength to deflect the charged particles and its atmosphere is too thin to stop the ions. (Earth's magnetic field is sufficient to deflect the solar wind.) Interaction of the solar wind with the lunar soils and rocks has enriched the surfaces of these materials in carbon, nitrogen, and hydrogen (fig. 2-6), and some mature soils may have attained an apparent solar wind carbon saturation level of about 200 ppm C. Conclusions regarding the role of solar wind are based only in part on the evidence cited above, namely, the direct correlation of the abundances of these biogenic elements in lunar soils with the surface area of the soils, their exposure histories, and with solar wind implanted rare gas abundances.

Data on the range of isotopic compositions for carbon in lunar samples (table 2-2) reflect several sources of the element, including terrestrial contamination. Values of $\delta^{13}C$ that fall within the range for terrestrial organic matter, –35 to –10 per mil, are highly suspect and indicative of major contributions from terrestrial contamination. Note, however, that bulk carbon isotopic compositions of meteorites lie typically in the range –25 to –5 per mil, making it difficult to distinguish isotopically between a

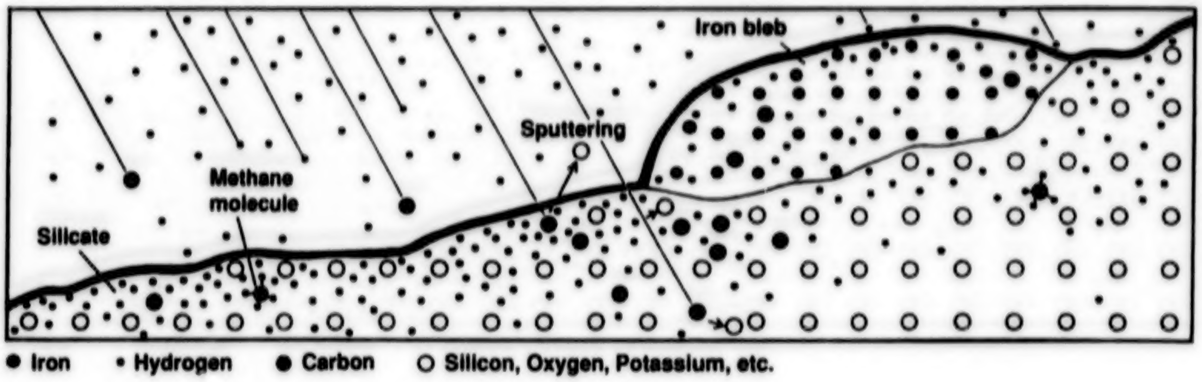


Figure 2-6. Solar wind plays an important role in the carbon chemistry of the lunar surface. An area of lunar soil is enlarged some 10^8 times to show a silicate grain with an iron bleb adhering to it. Carbon atoms implanted by the solar wind could be converted to methane by the vast excess of hydrogen in the solar wind. Damage to the crystalline structure of the silicate creates defects and voids that accommodate the gas and allow some compounds to escape by diffusion. Hydrocarbons containing two carbon atoms are probably generated as well with decreased efficiency. Carbon ions implanted into the iron bleb might react chemically to form the carbide-like species in lunar soils which on treatment with deuterated mineral acids yield deuterated hydrocarbons. Other elements in the solar wind, such as nitrogen, oxygen, and rare gases, are simultaneously implanted to a depth of a few hundred angstroms. (After Eglinton et al., 1972.)

meteoritic and terrestrial source. Positive values for soils, however, appear to reflect a non-terrestrial, non-meteoritic component of probable solar wind origin. The several ppm methane observed in lunar soils and believed to have been produced in soils by solar wind irradiation (see below) exhibited positive $\delta^{13}C$, including the highest (+46 per mil) found thus far in a naturally occurring organic compound. The very high positive $\delta^{13}C$ values in rocks coupled with their extremely low abundances reflect the production of carbon in their interiors by cosmic ray spallation reactions.

In conjunction with analyses of lunar samples, solar wind simulation experiments involving ion irradiation of lunar materials and terrestrial analogues were carried out. These simulation experiments together with sample analyses provide persuasive evidence of the role of solar wind in the chemistry of the biogenic elements on the Moon. Some experiments were designed to determine the molecular forms in which isotopically labeled ions (e.g., D^+, $^{13}C^+$, $^{15}N^+$) injected into samples at solar wind energies would be extracted and the temperatures over which such extraction occurred during vacuum pyrolysis. These experiments showed that carbon was evolved predominantly as ^{13}CO with minor amounts of $^{13}CO_2$, and nitrogen was evolved as $^{15}N_2$. Each of these labeled species exhibited essentially the same pattern of gas release over the 400-1300°C range as did the respective unlabelled gas that was derived from the indigenous carbon and nitrogen in the lunar soils. When extracted by vacuum pyrolysis, CO, CO_2, and N_2 accounted for all but a few percent of the total carbon and nitrogen in a lunar soil, a finding that is at least consistent with a solar wind source for most if not all of these elements.

In other experiments, serial irradiations with $^{13}C^+$ and D^+ ions led to production of ppm amounts of multiply labeled methane whose pattern of release from the artificially irradiated samples by heating under vacuum was the same as that of both the unlabeled methane extracted from lunar soils and the solar wind implanted noble gases. Apparently, solar wind irradiation can result not only in the retention of chemically bound species of carbon and nitrogen in the mineral matrix, but also the synthesis and trapping of the simplest of organic compounds, thus accounting for the methane and possibly other low molecular weight hydrocarbons found in lunar soils.

Table 2-2: Isotopic Compositions of Lunar Soils and Rocks(δ– Values in per mil)*

	Soils	Rocks
Hydrogen		
Range	–115 to –807	–10 to –97
Carbon		
Range	–35 to +46	–35 to +64
Nitrogen		
Range	–90 to +96	0 to +62
Sulfur		
Range	+6 to +11	–2 to +2.5

*For hydrogen,

$$\delta D = \frac{(D/H)_{\text{sample}} - (D/H)_{\text{standard}}}{(D/H)_{\text{standard}}} \times 10^3$$

Standard = standard mean ocean water

*For carbon,

$$\delta C = \frac{(^{13}C/^{12}C)_{\text{sample}} - (^{13}C/^{12}C)_{\text{standard}}}{(^{13}C/^{12}C)_{\text{standard}}} \times 10^3$$

Standard = Peedee belemite limestone

*For nitrogen,

$$\delta N = \frac{(^{15}N/^{14}N)_{\text{sample}} - (^{15}N/^{14}N)_{\text{standard}}}{(^{15}N/^{14}N)_{\text{standard}}} \times 10^3$$

Standard = air

*For sulfur,

$$\delta S = \frac{(^{34}S/^{32}S)_{\text{sample}} - (^{34}S/^{32}S)_{\text{standard}}}{(^{34}S/^{32}S)_{\text{standard}}} \times 10^3$$

Standard = Canyon Diablo meteorite troilite

The nature of the chemically bound carbon in the soils is not entirely understood. About 5 to 30% of this carbon can be released by deuterated mineral acids in the form of simple C_1 to C_3 hydrocarbons containing deuterium. This behavior is similar to that exhibited by some carbide minerals and by carbon dissolved in metallic iron. In these experiments with deuterated acids, the trapped solar wind synthesized methane is released in undeuterated form. The carbon of the deuterated hydrocarbons is typically enriched in ^{13}C relative to average terrestrial and meteoritic carbon and is characterized by positive $\delta^{13}C$ values

(+5 to +20 per mil) comparable to those of the uncontaminated bulk soils. This non-volatile, carbide-like carbon, like the bulk of the remaining uncharacterized carbon, is converted by pyrolysis (see above) to CO (predominantly) and CO_2 with isotopic compositions in the same range. The similarity in isotopic composition of the trapped gas, the carbide-like species and the remaining carbon, which may be trapped interstitially in minerallic, metallic, or glassy phases, argues strongly for a predominant contribution from a common source, that is, the solar wind.

The release of carbon mostly as CO rather than CO_2 during pyrolysis probably reflects the highly chemically reduced nature of the lunar soils due to the presence of metallic iron (fig. 2-6). Unlike terrestrial basalts, lunar basalts also contain metallic iron and are, therefore, highly reduced. If any carbon-containing magmatic gases accompanied the lavas that flowed across the Moon's surface, they too would have been dominated by CO. In contrast, CO_2 is predominant in terrestrial volcanism. If the Moon ever had a thick atmosphere produced by global outgassing, the reduced chemical composition of the atmosphere would have been conducive to the synthesis of organic matter. This conclusion is based on the highly successful results of Miller-Urey type experiments designed to simulate chemical evolution in Earth's prebiotic atmosphere. No evidence of such synthesis has been found in lunar samples.

Instead, any organic chemical synthesis or high pressure hydrogenation reaction that occurs on the Moon results from ion-molecule chemistry driven by the interaction of solar wind ions with mineral grains, and only simple hydrocarbons are produced within a thin surface layer of the grains. Such synthesis could also occur on the surfaces of bodies in the asteroid belt and satellites of the outer planets. A related process could also be operating in the upper atmosphere of Titan where solar wind interactions with gases could lead to synthesis of organic compounds. Moreover, the organic chemistry of interstellar clouds is thought to arise from ion-molecule reactions driven by cosmic ray interactions with gas and dust. Thus, the limited organic synthesis occurring on the Moon is but one manifestation of a type of physical-chemical process that occurs throughout the cosmos.

Although the effects of another widely occurring process, bombardment of planetary surfaces by meteorites and comets, have not left a clearly discernible imprint in carbon and other biogenic elements on the Moon (see below for sulfur and phosphorus), there are several reports of lunar samples containing small amounts of water and other volatiles which might be interpreted as vestiges of impacting objects. For example, several rocks from Apollo 16 contained the minerals goethite and lawrencite typically associated with rust. The formation of these minerals from water in terrestrial laboratories was also possible. One of the immature soils collected from the rim of Flag crater at the Apollo 16 site and sieved in the LRL to remove coarser than mm-sized particles was found to contain an unusual amount (up to 100 ppm) of HCN, CH_4, and H_2O. It was suggested that the soil contains the remnants of volatiles associated with a cometary impact. Such high abundances of volatiles, however, were not found in an unsieved sample of the same soil. These findings leave unresolved the question of direct evidence for cometary volatiles in lunar soils, but they do raise another interesting question. The time interval of cosmic ray exposure of the soils at the Apollo 16 site, where the putative "cometary components" were found, is similar in age to that of the K-T boundary on the Earth. Could this possibly suggest that on the Moon a record of cometary impacts might be found to correlate with those hypothesized to have been responsible for the extinction of the dinosaurs and other biota 65 million years ago?

The average nitrogen content found in 85 lunar soils was 82 ppm with a range of 7 to 164 ppm; immature soils contain less than 40 ppm N (table 2-1). Lunar breccias have an average of 68 ppm N with a range of values between 16 and 181 ppm N. Lunar rocks averaged 3 ppm N and ranged between 1 and 6 ppm. Early workers suggested that nitrogen abundances for lunar rocks were a factor of 10 greater but it has since been shown that the higher abundances resulted from terrestrial contamination of the samples, despite the precautions taken. Nitrogen contents of lunar soils are proportional to the carbon contents of the same soil. As for carbon, the data for nitrogen indicate that most of the nitrogen in lunar soils originated in the solar wind. The total nitrogen content of lunar soils was found to be proportional to the solar wind ^{36}Ar and H. It was shown that the grain size separates of lunar fines indicate that the highest concentrations of nitrogen reside in the particles with the greatest total surface area (per unit mass) or smallest grain sizes. Because the retention of solar wind nitrogen in lunar materials was found to be highest of any solar wind component, the nitrogen content has become a useful indicator of soil maturity.

John Kerridge of the University of California at Los Angeles showed that the solar wind nitrogen, implanted in lunar soils, exhibits isotopic variations that are related to the time, although not to the duration, of implantation, with earlier samples characterized by lower ratios of ^{15}N to ^{14}N (table 2-2). The increase in the solar ^{15}N content of lunar materials during the lifetime of the lunar regolith was probably caused by spallation of ^{16}O in the surface of the Sun (fig. 2-7). From the study of the isotopic composition of the nitrogen present within the lunar regolith, the early history of the Sun might be tracked.

The hydrogen distributions in lunar materials are similar to those of carbon and nitrogen, and are predominantly attributable to a solar wind origin. The isotopic composition of the hydrogen (table 2-2) shows a depletion of deuterium in the lunar materials consistent with a solar source. The finite abundance of deuterium, however, must be attributed to terrestrial contamination or a meteoritic contribution; the solar wind is free from deuterium and thus would not contribute to its abundance. Hydrogen abundances are directly related to length of time of surface exposure to the solar wind. The greatest hydrogen abundances are found in the finest grain size fractions of the soils. For example, the sub-45 micrometer grain size fraction of a mature lunar soil contains between 40 and 145 ppm hydrogen. Hydrogen concentrations for bulk lunar soils range from 3.2 to 60 ppm. Mature lunar soils typically contain greater than 50 ppm H. The lowest hydrogen concentrations are found for immature lunar soils. The hydrogen abundances are correlated with the soil maturity index, I_s/FeO (an indicator of the fine-grained

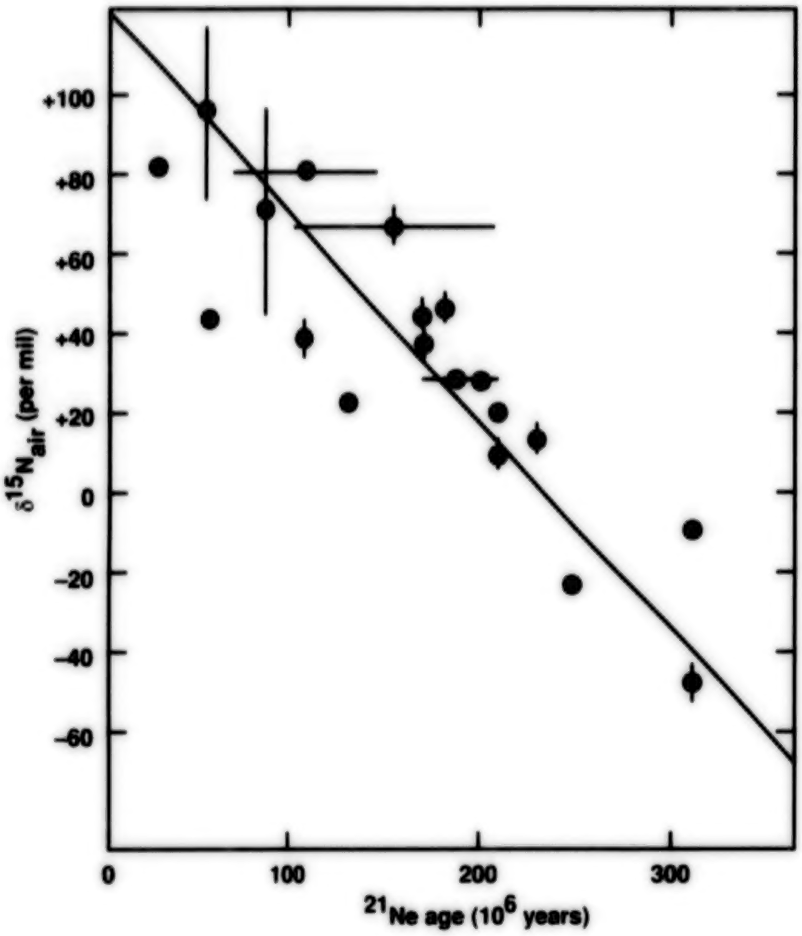


Figure 2-7. The isotopic ratio of ^{15}N to ^{14}N (represented by $\delta^{15}N$) values for Apollo 16 soil samples show a strong negative correlation with the cosmic-ray exposure ages. Kerridge has suggested that the solar abundance of ^{15}N has increased during the history of the Sun. In the solar interior ^{15}N is burned before ^{14}N. The observation that the abundance of ^{15}N has increased with time in the surface of the Sun shows that convective mixing involving the deep interior of the Sun has been insignificant. In addition, the apparent lack of lunar fractionation in nitrogen is consistent with highly efficient retention of this element on the lunar surface.

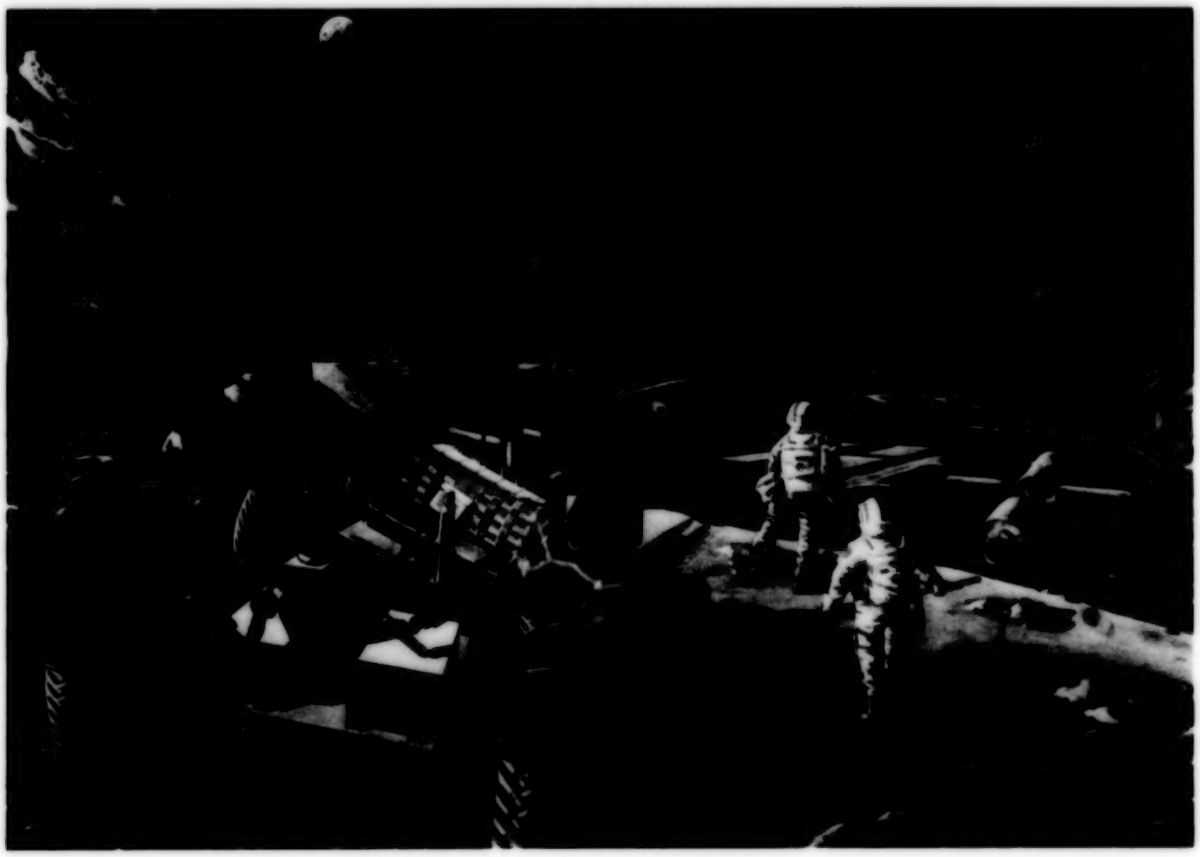
metallic iron present in the soil which results from micrometeorite impact processes). Supposedly, the heat released during impacts causes the solar wind hydrogen in the soil to reduce small amounts of ferrous iron in the minerals to metallic iron.

Understanding the hydrogen abundances and distributions in lunar materials is important if hydrogen is to be used as a consumable or propellant at a future lunar base (fig. 2-8). Hydrogen is also required for the ilmenite ($FeTiO_3$) reduction process proposed for the production of oxygen at a lunar base. Water, the by-product of the reduction process, can be electrolyzed to produce both oxygen for use in the habitats and hydrogen for use again to reduce additional ilmenite.

Total sulfur abundance measurements along with isotopic ratio measurements of lunar soils have shown significant variations between lunar soils, breccias, and rocks. Lunar soils from Apollo 11, 12, 15, and 17 are depleted in their total sulfur content relative to the mare basalts found at each landing site. The dominant form of sulfur in lunar materials is the mineral troilite, FeS, although trace quantities of other metal sulfides have been reported. The total sulfur content of lunar soils from all of the Apollo missions has been accounted for by mixing models based upon the sulfur contribution from individual components found in the lunar fines at a particular sample site. Addition of sulfur to the lunar regolith from meteorite infall is not needed to account for the observed sulfur abundances. Carbonaceous chondrite meteorites (Type I) contain up to 6% total sulfur. For any lunar mixing model, large amounts of sulfur cannot have been added to the lunar surface from meteorites (or the total sulfur content of the soils would be too much to be accounted for by the mixing models) unless some major loss mechanism for sulfur has operated.

The average total sulfur content of lunar soils is 850 ppm (192 samples analyzed) with a range of between 290-1400 ppm. Lunar highland soils have the lowest total sulfur abundances (between 290 and 600 ppm). The isotopic composition of the sulfur (table 2-2) associated with lunar soils is isotopically heavy (ranging from +6 to +11 per mil) relative to the meteorite troilite standard. The enrichment has been attributed to the loss of the isotopically light sulfur phases present in the surface materials during reaction with the solar wind hydrogen to form H_2S during micrometeorite bombardment and surface cratering processes. The mean sulfur content of 96 lunar breccias is 756 ppm with a range of 20 to 2250 ppm. Lunar crystalline rocks range in sulfur content between 20 and 2800 ppm (92 samples) with a mean value of 1085 ppm. Anorthosite samples from the lunar highlands or early lunar crust contain the least amounts of sulfur (below 300 ppm) while the mare basalts from Apollo 11 and 17 contain the greatest amounts of sulfur (mean values of 2033 and 1860 ppm, respectively). The sulfur contents of the lunar basalts

Figure 2-8. Surface operations at a lunar base. Two future lunar astronauts are shown overlooking the lunar base habitat and surface mining operations. (NASA-JSC Photograph S86-27256, Artwork by Dennis Davidson.)



are almost 10 times greater than values seen from terrestrial subaerial basalts. A comparison of the sulfur abundances of lunar basalts with submarine basalts shows that lunar samples are enriched two to four times those of terrestrial basalts which contain the greatest amounts of sulfur. This suggests that the source regions where the lunar basalts were formed contained greater amounts of sulfur than those of terrestrial basalts. The isotopic composition of the sulfur associated with lunar crystalline rocks is remarkably constant (ranging from –2 to +2.5 per mil) and similar to the reference meteoritic troilite standard (table 2-2) in contrast to other objects within our solar system.

Phosphorus is present in all lunar samples in trace amounts. For most lunar basalts the range of P contents is quite limited, being about 200 to 900 ppm. The major phosphorous-bearing phase is apatite, $Ca_5(PO_4)_3(F,Cl)$, although whitlockite, $Ca_3(PO_4)_2$, is also present along with traces of meteoritic schreibersite $(Fe,Ni)_3P$. Lunar highland samples may contain the phosphorous-bearing KREEP (potassium, rare Earth elements, and phosphorous) glass which contains a relatively high phosphorous content of around 0.33% by weight.

The experience with lunar samples has shown clearly the great scientific insight that can be gained by making high-precision measurements on Earth of material returned from other solar system bodies. The variety of techniques which can be used for sample preparation and analysis has increased immensely, and studies in terrestrial laboratories offer the maximum flexibility for designing new experiments or modifying sample preparation to clarify ambiguous results and optimize scientific return. Moreover, as was the case for lunar samples, new analytical techniques developed after the return of extraterrestrial samples can be brought into play. Clearly, study of samples from Mars or comets in terrestrial laboratories will afford data far superior in quality and quantity to that obtainable by any remote analysis on a distant planet.

After nearly two decades of studying the lunar samples some major questions remain which are related to the history and evolution of the biogenic elements during lunar history. Briefly, some of the important questions are: Can the severe depletion of biogenic elements and other volatiles in the Moon be accounted for by an impact origin for the Moon? What is the exact composition of the volcanic gases and volatiles, if any were associated with the lunar magmas which poured out onto the lunar basins? What are the volatiles associated with the young volcanic vents observed in several regions of the lunar maria basins? Are there sinks for the lunar volatiles at the poles and permanently shadowed regions of the Moon? What are the concentrations of OH^-, H_2O, and methane associated with the surfaces of lunar grains resulting from the solar wind irradiation of the lunar surface? What are the production and loss rates of the simple molecules such as methane produced by solar wind irradiation? Are there regions on the lunar surface where the remnants of cometary impacts can be observed and studied? Is there a record of periodic bombardment of the lunar surface by comets or asteroids that might be correlated with similar processes on Earth? How can the lunar regolith or soil be converted to a medium which will support plant growth or other food production processes in association with a lunar base? Many of these questions may best be addressed by further exploration of the Moon, including the return of materials from previously unsampled regions and the conduct of experiments on the lunar surface.

The miniscule amount of organic matter in lunar surface materials and the absence of any life forms are consistent with findings that the lunar surface has undergone melting events and eons of solar and cosmic radiation and meteorite impacts under conditions of high vacuum. These processes appear to have obliterated all traces of any primitive organic synthesis on the lunar surface. Yet, carbon and carbon-containing compounds indeed are found in the samples. They signify that chemical evolution, though severely constrained by the lack of atmosphere and the depletion of volatiles on the moon, has in fact proceeded to a very limited extent and is due virtually entirely to solar wind interactions with the lunar surface. Thus, the Apollo and Luna missions have afforded unique oppoprtunities to study aspects of the chemistry of carbon and the other biogenic elements which have no natural counterparts on Earth. The analyses of lunar samples by organic and analytical chemists became not only searches for compounds related to terrestrial biochemicals, but also primarily studies of lunar chemical evolution and fundamental aspects of the cosmo-chemistry of the biogenic elements. This is only fitting if one realizes that the Sun has been the dominant factor affecting the chemistry of the biogenic elements on the Moon and that in the Sun hydrogen is the most abundant element while carbon and nitrogen are the fourth and fifth most abundant elements.

Additional Reading

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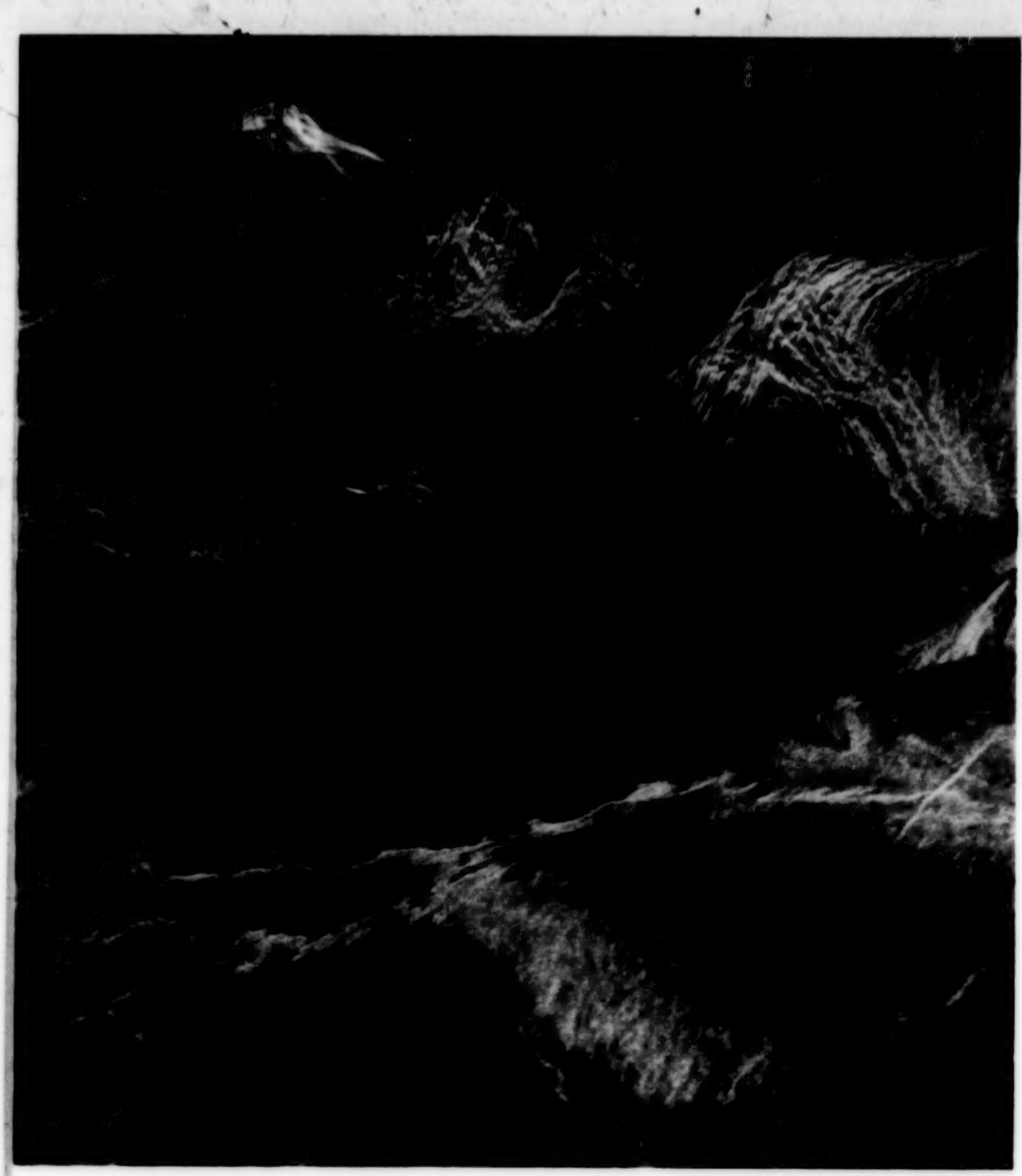

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Venus:
A Search for Clues to Early Biological Possibilities

L. Colin and J. F. Kasting

Extant life on Venus is out of the question. The current atmospheric environment at the surface of the planet is far too hostile, by benign terrestrial standards, to support life or to participate in the origin of life. In the first part of this paper, we will summarize the extensive evidence for this assertion.

If the assertion is correct, then why are exobiologists at all interested in Venus? One answer to this question involves the possibility of *extinct* life. Although few scientists would consider this to be very likely, life may have had some chance to originate on Venus because Venus may have been considerably more Earth-like in its past. A second answer is that the study of Venus may teach us something about life on Earth, even if Venus itself has always been lifeless. In particular, Venus may tell us what the physical limits are on the habitability of

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an Earth-like planet and how likely it is that other habitable planets exist elsewhere in the galaxy.

In the second part of this paper, we will address these fundamental questions. But before we do so, we must discuss the current state of the planet and its environment. Knowledge of origin and evolution can only be inferred using the knowledge of the present state of the planet along with appropriate theory and modeling.

It should be noted that for Mars, the possibility of *extant* life has not been ruled out, even by the Viking missions. Future missions to the red planet will thus be instrumented to search for signs of both *extant* and *extinct* life, although the latter is being given highest priority. For Venus, only *extinct* life is feasible, but the environment is so hostile that it will be very difficult, and probably impossible with current technology, to even plan for such a mission in this century.

Current State of Venus and its Environment

Venus has been the target of more unmanned, scientifically instrumented, interplanetary spacecraft than any other object in the solar system (table 3-1). The Soviet Union, until recently when it announced its intention to switch its focus to Mars for the remainder of the century, at least, has historically concentrated its solar system exploration program on Venus. From 1961 through 1985 they sent Venera and Vega planetary flybys, planetary orbiters, atmospheric entry probes/surface landers, and balloons to our nearest neighbor.

The United States, although spreading its resources on spacecraft missions designed to visit most of the planets of the solar system in a more "balanced" program, also launched Mariner and Pioneer planetary flybys, planetary orbiters, and atmospheric entry probes to Venus from 1962 to 1978. The Pioneer Venus Orbiter, after 13 years in orbit, continues to collect and telemeter valuable data back to Earth. Magellan, an orbiter dedicated to high resolution surface mapping and gravity observations of Venus, was launched in 1989. It is currently on an extended phase after completing its highly successful nominal mission. The result of these many, sophisticated missions is a set of data and understanding unmatched by any planet except our own. In this part of the paper, we summarize only that data and understanding critical to the questions posed by exobiologists.

When one looks at those critical characteristics of a planetary environment most pertinent to exobiology, one is immediately struck by the extreme values associated with Venus (table 3-2). For example, the average surface temperature of Venus is 464°C (867°F)—about twice as hot as the maximum setting on a kitchen oven! Furthermore, the surface temperature does not vary from this figure by more than a few degrees centigrade from noon to midnight, or from equator to pole, perhaps 5-15°C. This contrasts sharply with an average surface temperature of 15°C (59°F) on Earth, with noon-midnight average differences of 10°C and equator-pole average differences of 45°C.

Table 3-1: Planetary Spacecraft Missions to Venus

Name	Launch date	Remarks
Venera 1	2/12/61	Attempted flyby. Contact lost 2/27/61. Estimated to have passed within 100,000 km of Venus and continued into heliocentric orbit. Considered failure.
Mariner 1	7/22/62	Attempted flyby. Booster destroyed by ground control after 5 min of flight. Considered failure.
Mariner 2	8/27/62	Flyby 12/14/62 at 34,833 km closest approach. First successful probe to another planet.
Venera 2	11/12/65	Flyby 2/27/66 at 24,000 km closest approach. Communication failed just before flyby. Considered failure.
Venera 3	11/16/65	Atmospheric entry probe. Communication link failed just before entry 3/1/66. Considered failure.
Venera 4	6/12/67	Atmosphere entry probe. Entered 10/18/67. Radio transmitter failed at 27 km altitude. First successful USSR mission.
Mariner 5	6/14/67	Flyby 10/19/67 at 3391 km closest approach.
Venera 5	1/5/69	Atmospheric entry probe/soft lander. Entered 5/16/69. Radio signals from probe ceased at 25 km altitude.
Venera 6	1/10/69	Atmospheric entry probe/soft lander. Entered 5/17/69. Radio signals from probe ceased at 11 km.
Venera 7	8/17/70	Atmospheric entry probe/soft lander. Entered 12/15/70. Transmitted onsurface for 23 min.
Venera 8	3/26/72	Atmospheric entry probe/soft lander. Entered 7/22/72. Transmitted on surface for 50 min.
Mariner 10	11/3/73	Flyby 2/5/74 at 5793 km closest approach.
Venera 9	6/8/75	Combined orbiter and atmospheric entry probe/soft lander. Orbit insertion and entry 10/22/75. Transmitted on surface for 53 min.
Venera 10	6/14/75	Combined orbiter and atmospheric entry probe/soft lander. Orbit insertion and entry 10/25/75. Transmitted on surface for 65 min.
Pioneer Venus 1	5/20/78	Orbiter. Inserted 12/4/78. Spacecraft still functional.
Pioneer Venus 2	8/8/78	Multiple atmospheric entry probes (4) plus upper atmosphere probe (Probe Bus). Entered 12/9/78.
Venera 11	9/9/78	Combined flyby and atmospheric entry probe/soft lander. Entered 12/21/78. Transmitted on surface for 95 min. Flyby 12/21/78 at 25,000 km closest approach.
Venera 12	9/14/78	Combined flyby and atmospheric entry probe/soft lander. Entered 12/25/78. Transmitted on surface for 110 min. Flyby 12/25/78 at 25,000 km closest approach.
Venera 13	10/30/81	Combined flyby and atmospheric entry probe/soft lander. Entered 3/2/82. Transmitted on surface for 127 min.
Venera 14	11/4/81	Combined flyby and atmospheric entry probe/soft lander. Entered 3/5/82. Transmitted on surface for 53 min.
Venera 15	6/2/83	Orbiter Radar Mapper. Inserted 10/10/83.
Venera 16	6/7/83	Orbiter Radar Mapper. Inserted 10/14/83.
Vega 1	12/15/84	Combined balloon and atmospheric entry probe/soft lander. Entered 6/10/85. Balloon operated at 55 km altitude for 2 days. Transmitted on surface for 56 min.
Vega 2	12/21/84	Combined balloon and atmospheric entry probe/soft lander. Entered 6/15/85. Balloon operated at 55 km altitude for 2 days. Transmitted on surface for 57 min.
Magellan	1989	Orbiter Radar Mapper. Spacecraft still functional.

Table 3-2: Venus-Earth Comparisons

Parameter	Earth	Venus
Mass (Earth = 1)	1.000	0.815
Mean radius, km	6378	6051.5
Oblateness	0.003	0
Mean planet density, gm cm^{-3}	5.52	5.24
Surface gravity (Earth = 1)	1.00	0.88
Escape velocity, km sec^{-1}	11.2	10.4
Mean solar distance, AU	1.000	0.723
Solar constant, kw m^{-2}	1.38	2.62
Solar revolution period, days	365.26	224.7
Rotational period	$23^h56^m23^s$ E	243.01 days W
Sol-Earth days	1	117
Orbital eccentricity	0.017	0.007
Inclination to orbit plane, deg	23.45	177.4*
Orbit inclination to ecliptic, deg	3.394	0.000
Magnetic moment, gauss cm^{-3}	7.91×10^{25}	$<10^{22}$
Bond albedo	0.30	0.77
Effective temperature, K	255	229
Average surface temperature, K	288	737
Greenhouse magnitude, K	33	508
Mean surface pressure, bars	1.013	95
Atmosphere/planet mass	8.8×10^{-7}	9.81×10^{-5}
Total surface relief, km	20	13

*Inclinations >90° imply retrograde rotation.

It is tempting to explain the difference in average surface temperatures between Venus and Earth by the fact that Venus is much closer to the Sun. This fact is, of course, true and, furthermore, it is certainly true that solar radiation is overwhelmingly the most important energy source for heating the terrestrial atmospheres. Internal heat sources are important only for the outer, giant gas planets.

The solar flux at the mean orbital distance of Venus is some 1.9 times that at Earth, since Venus' mean distance from the Sun is 0.72 astronomical units (AU) compared with the Earth value of 1.00 AU (fig. 3-1). However, the Bond albedo of Venus is significantly higher than that of Earth, 0.77 versus 0.30, so that a much larger fraction of the incident solar flux is reflected back into space. The net result is that Venus absorbs almost 40% less energy than does Earth, or only slightly more energy than is absorbed by Mars.

Venus' albedo is larger than that of Earth primarily because of the ubiquitous nature of its clouds. Venus is 100% cloud covered at all times, whereas Earth is about 50% cloud covered at any time. On Venus, most of the absorbed solar radiation occurs in the clouds, and only a small percentage reaches the ground. On Earth, most of the incident solar radiation is absorbed at the ground.

All of this suggests that the surface temperature of Venus should be colder than the surface temperature at Earth, not hotter. If the amount of absorbed solar energy were the only important factor, Venus and Earth would have surface temperatures of –44°C and –18°C, respectively.

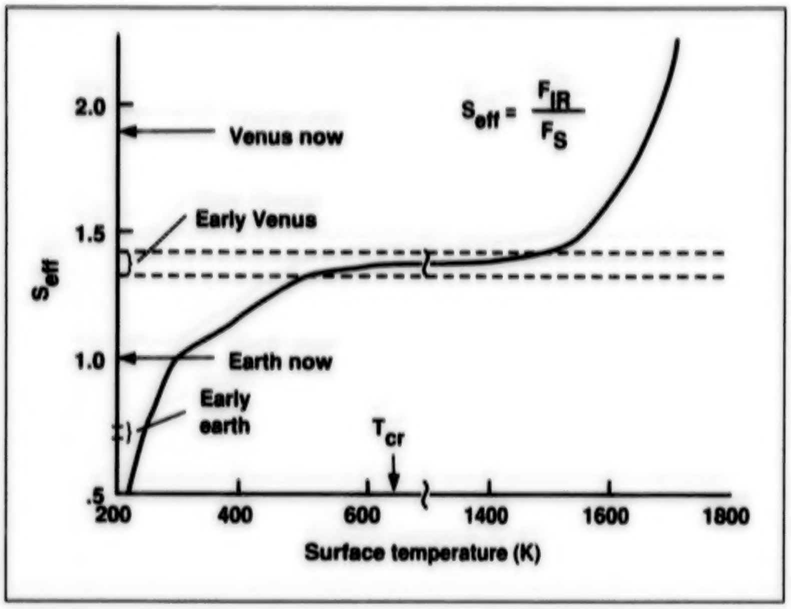


Figure 3-1. Effective solar constant S_{eff}. The horizontal dashed lines represent estimates of the solar flux at Venus orbit 4.6 billion years ago. There is a break in the horizontal scale between 700 and 1300 K.

Clearly, something has been left out of our argument—the Greenhouse Effect. The Greenhouse Effect has become a household word in recent years as the Earth's atmosphere has been warmed by the injection of carbon dioxide and other pollutants. Briefly, the way this effect works is as follows. The solar radiation is spread out over a fairly broad fraction of the electromagnetic spectrum—from the ultraviolet through the visible into the infrared. However, the bulk of the energy is concentrated in the visible, where the atmosphere of Earth, and to a lesser extent the atmosphere of Venus, is largely transparent. Thus, visible radiation reaches the surface and heats it. The heated surface reradiates this energy as heat, or infrared radiation. Certain atmospheric gases, notably carbon dioxide and water vapor, absorb strongly in the infrared, rendering the atmosphere partially opaque at those wavelengths. Thus, the atmosphere itself becomes heated, and radiates in the infrared. Some of this energy escapes to space and some of it heats the surface still further.

Eventually, a balance is struck between the incoming solar energy and the outgoing infrared energy resulting in a stable temperature at the surface. The average temperature of the Earth is 15°C, or some 33°C higher than it would have been without the Greenhouse Effect. For Venus, the average temperature is 464°C, 508°C of which is caused by greenhouse warming. The much larger Greenhouse Effect on Venus is a consequence of its dense carbon dioxide atmosphere (fig. 3-2).

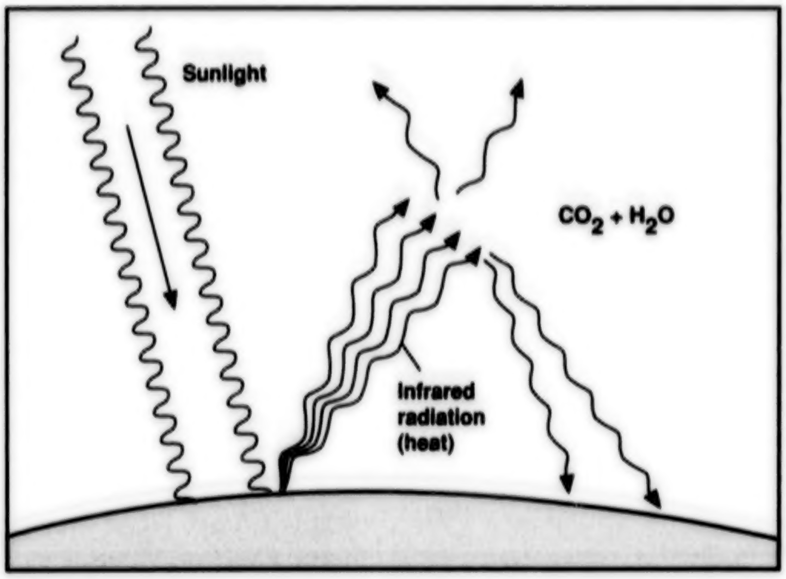


Figure 3-2. Greenhouse Effect occurs when certain gases, notably carbon dioxide and water vapor, warm the surface of a planet. Such gases allow light from the sun to reach the planet, but they intercept the infrared rays (heat) that the planet radiates into space and reradiate much of this energy toward the surface. The gases raise the Earth's surface temperature some 35°C above what it would be if they were absent.

Another interesting difference between Venus and Earth concerns short-term (diurnal) temperature variations. It is important to recognize that a day on Venus is much different than on Earth. Because of Venus' comparable orbital period (about the Sun)—224.7 days—and spin period (about its axis, retrograde)—243 days—Venus has a "day" equal to 117 Earth days. Thus, the observation that the noon and midnight surfaces differ in temperature by only a few degrees is even more impressive. There would appear to be plenty of time for a region in the long night sector to cool off, by radiating its heat to space. Since this is not the case, we must conclude that the flow of heat around the planet is efficient and very rapid. A more precise way of stating this is to say that the time constant for heat transport is much shorter than the time constant for radiative transport. The radiative time constant at the surface of Venus, that is the heat capacity of the atmosphere divided by the outgoing infrared flux, is about 127 Earth years, much longer than the lengthy Venus day. The equivalent time constant for Earth is about four months.

Seasonal variations are also much smaller on Venus. One reason is that Venus has a very small orbital eccentricity about the Sun (0.007) compared with a value of 0.017 for Earth. Thus, the distance between Venus and the Sun varies only slightly from "summer" to "winter." Venus' obliquity, that is, the angle between its spin axis and the normal to its orbital plane, is also very small, 2.5°, compared to 23.5° for the Earth. Thus the northern and southern hemispheres receive essentially equal amounts of radiation year round.

The cause of these extreme temperature conditions on Venus is its extremely massive atmosphere. At the mean planet radius, the atmospheric pressure is 95 bars ($1.013 \text{ bar} = 1 \text{ atm} = 1.013 \times 10^6 \text{ dynes cm}^{-2}$). Since the mean molecular weight is 43.44, the atmospheric column density is about 65 kg m^{-3}. Thus the Venus atmosphere is nearly 100 times denser than that of Earth.

The Venus atmosphere near the surface is composed of two major gases: carbon dioxide (CO_2) and molecular nitrogen (N_2). The volume mixing ratios of these two gases are about 96.5% and 3.5%, respectively. Many other gases have been measured in trace amounts and others are suspected. For many of the measured trace constituents there is significant controversy as to their precise relative abundances.

It may seem surprising to the uninitiated that CO_2 is the major gas in the Venus atmosphere given that it is only a minor constituent (0.034%) in the Earth's atmosphere. It is, of course, an important trace constituent to the life cycle because of the role it plays in photosynthesis. Most of the CO_2 on Earth, however, is locked up as (calcium and magnesium) carbonates in rocks. These carbonate rocks are formed by reactions that take place in liquid water. The lack of an ocean and thus a hydrological cycle on Venus has allowed nearly all the planet's CO_2 to remain in the atmosphere. The total inventory of CO_2 on each planet is, in fact, practically the same. As on Earth, other trace atmospheric constituents play important roles in the physics and chemistry of the planet, e.g., as catalysts in chemical and photochemical processes, in meteorological processes, and in atmosphere-surface interactions. Unlike Earth, these processes are strictly abiotic on Venus. On the Earth, the two major gases, molecular nitrogen (N_2) and molecular oxygen (O_2), and many of the trace constituents have biological processes as their major sources and major sinks. On Venus, the source of N_2 is outgassing from the interior of the planet, and sinks are non-existent. Attempts at measuring O_2 have not been successful. The important role that life plays in the chemical make-up of Earth's atmosphere makes any comparison of the atmospheres of Venus and Earth almost moot. Earth's atmosphere would not be anything like it is today, had life not formed and flourished.

Another trace gas (with a maximum mixing ratio of 1×10^{-7}) on Earth of critical importance to the survival of life is ozone (O_3). Found mainly in the stratosphere, it effectively absorbs solar ultraviolet radiation that would be lethal to life. Scientists have so far been unable to identify any ozone at Venus, but have been able to set an upper limit to its mixing ratio of 1×10^{-6} to 1×10^{-7}.

From an exobiologist's viewpoint, liquid water on the surface and water vapor (H_2O) in the atmosphere are essential. Liquid water cannot exist on the surface of Venus because the surface temperature exceeds the critical temperature of water, 374°C. (The critical temperature is the highest temperature at which the liquid phase can exist.) Thus, any liquid water that may once have existed on the surface of Venus would have evaporated into the atmosphere. However, today we find very little water vapor in the atmosphere. The precise volume mixing ratio is controversial, but it appears to be 2×10^{-4} or less in the lower atmosphere. The variable mixing ratio of water vapor in the Earth's atmosphere is about 4×10^{-2} or less; here, of course, atmospheric water vapor is in contact with a much larger water reservoir—the oceans.

Whether Venus formed with liquid water on its surface, acquired it somehow in its early history and lost it through evaporation, is unknown and controversial. It is the subject of the second part of this paper. But if it did have an ocean and lost it through evaporation to the atmosphere, then where did the water vapor go?

One likely possibility is that the vapor was photolyzed by solar ultraviolet radiation into hydrogen and oxygen. One might then expect to find these remnants in the atmosphere. However, current measurements suggest only trace amounts of each, perhaps 0.002% to 0.003% by volume. If these gases are not in the atmosphere, where else might they be?

The hydrogen molecules may have been photolyzed into hydrogen atoms which, being light, may have escaped Venus' gravity into space. The oxygen atoms from the photolyzed water vapor may have reacted with surface materials to become locked in the regolith and lithosphere. At any rate, the question of water on Venus in its earlier history is still open.

Another non-biological reason for the importance of water is its role as an infrared-active gas in the "Runaway Greenhouse Effect." Recall that we postponed distinguishing it, with its apparent important part it plays at Venus, from the "ordinary" Greenhouse Effect operative for Earth. The following "idea experiment" nicely illustrates the Runaway Greenhouse Effect.

Suppose we were to move Earth from its current position (1.00 AU) to the position of Venus (0.72 AU). The consequences would be

1. Oceans become warmer;

2. More water evaporates;

3. Increased water vapor in the atmosphere blocks infrared radiation from the surface, thereby increasing the surface temperature.

The cycle would repeat until

1. Oceans boil away, the atmosphere becomes very hot and full of water vapor, which rises into upper levels of the atmosphere;

2. At high altitudes, ultraviolet light breaks water molecules into hydrogen and oxygen;

3. Hydrogen escapes from the planet, oxygen remains to combine with rocks, atmosphere becomes dry and full of CO_2. (Carbonates cannot form without liquid water, CO_2 is continually added to the atmosphere by volcanoes.);

4. Earth resembles Venus!

So the Runaway Greenhouse Effect requires an ever-increasing amount of evaporated water in the atmosphere from the oceans to cause the atmospheric temperature to rise to exceptional levels. CO_2 cannot do this alone; there are too many infrared leaks to space. The above scenario nicely illustrates what would happen to Earth if it were moved to Venus' orbit position. Whether it describes how Venus' current environment evolved depends on whether Venus had a sizable ocean earlier in its history.

Table 3-3: Composition of the Venus Troposphere (from Prinn and Fegley)

Gas	Volume mixing ratio	Major source	Major sink
CO_2	9.63×10^{-1}	Outgassing	$CaCO_3$ formation?
N_2	3.5×10^{-2}	Outgassing	—
CO	2×10^{-5} (22 km), 10^{-3} (100 km)	Photochemistry (CO_2)	Photooxidation
SO_2	1.5×10^{-4} (22 km), 5×10^{-8} (70 km)	Photochemistry	$CaSO_4$ formation
^{36}Ar	3.7×10^{-5}	Outgassing–(primordial)	—
^{38}Ar	3.7×10^{-5}		—
^{40}Ar	3.3×10^{-5}	Outgassing (40 K)	—
H_2O	10^{-4} (22 km), $(1\text{-}40) \times 10^{-6}$ (70 km) *	Outgassing, impacts	Silicate hydration, Fe^{++} oxidation H escape
H_2	$<2.5 \times 10^{-5}$*	Photochemistry	Escape as H
^{4}He	1.2×10^{-5}	Outgassing (U,Th)	Slow escape
H_2S	$(3\text{-}40) \times 10^{-6}$*	Outgassing (FeS_2)	Photooxidation
COS	$<4 \times 10^{-5}$*	Outgassing (FeS_2)	Photooxidation
^{20}Ne	7×10^{-6}	Outgassing–(primordial)	—
^{22}Ne	7×10^{-6}		—
^{80}Kr	7×10^{-7}*	Outgassing–(primordial)	—
^{82}Kr	7×10^{-7}*		—
^{84}Kr	5×10^{-8}*	Outgassing, ^{235}U	—
^{86}Kr	5×10^{-8}*		—
HCl	4×10^{-7}	Outgassing (NaCl)	NaCl formation
HF	5×10^{-9}	Outgassing (CaF_2)	CaF_2 formation

*Important disagreements exist between the different instruments that have measured these species.

Many other gases have been detected on Venus, but we have discussed those that are most closely connected with exobiology. Table 3-3 lists the volume mixing ratios of all of these, including their major sources and sinks. A similar table is reproduced for Earth (table 3-4).

The ubiquitous clouds that veil Venus are found at high altitudes, 50 to 70 km, above the surface, but at approximately the same pressure and temperature levels as on Earth. Unlike the water condensation clouds on Earth, the clouds of Venus are primarily composed of aqueous solutions of sulfuric acid (H_2SO_4). This sulfuric acid is produced from the photolysis of SO_2 that diffuses up from the lower atmosphere. As discussed earlier, the clouds absorb or reflect most of the incident solar radiation allowing only a small amount to reach the surface. Another potential surface effect is acid rain; there were some indications on earlier Venera

Table 3-4: Composition of the Earth Troposphere
(from Prinn and Fegley)

Gas	Volume mixing ratio	Major source	Major sink
N_2	7.81×10^{-1}*	Biology	Biology
O_2	2.09×10^{-1}*	Biology	Biology
^{40}Ar	9.3×10^{-3}*	Outgassing (^{40}K)	—
H_2O	$<4 \times 10^{-2}$	Evaporation	Condensation
CO_2	3.4×10^{-4}	Combust., biology	Biology
^{36}Ar	3.7×10^{-5}	Outgassing– (primordial)	—
^{38}Ar	3.7×10^{-5}		—
^{20}Ne	1.82×10^{-5}	Outgassing– (primordial)	—
^{22}Ne	1.82×10^{-5}		—
^{4}He	5.24×10^{-6}	Outgassing (U, Th)	Escape
CH_4	$1.7\text{-}3 \times 10^{-5}$	Biology	Photooxidation
^{80}Kr	1.14×10^{-6}	Outgassing (^{235}U)	—
^{82}Kr	1.14×10^{-6}	Outgassing (^{235}U)	—
^{84}Kr	1.14×10^{-6}	Outgassing (^{235}U)	—
^{86}Kr	1.14×10^{-6}	Outgassing (^{235}U)	—
H_2	5×10^{-7}	Photochem. (H_2O)	Escape as H
N_2O	3.1×10^{-7}	Biology	Photodissociation
C_2H_4, etc.	$<7 \times 10^{-7}$	Incomplete comb.	Photooxidation
C_2H_2, etc.	$<2 \times 10^{-7}$	Incomplete comb.	Photooxidation
C_4H_{10}, etc.	$<2 \times 10^{-7}$	Incomplete comb.	Photooxidation
Toluene, etc.	$<1 \times 10^{-7}$	Incomplete comb.	Photooxidation
CO	$(0.4\text{-}2) \times 10^{-7}$	Photochemistry	Photochemistry
^{128}Xe	8.7×10^{-8}	Outgassing (U, I)	—
^{132}Xe	8.7×10^{-8}	Outgassing (U, I)	—
^{134}Xe	8.7×10^{-8}	Outgassing (U, I)	—
^{136}Xe	8.7×10^{-8}	Outgassing (U, I)	—
O_3	$(0.1\text{-}1) \times 10^{-7}$	Photochem. (NO_2)	Photochemistry
CH_3O_2H, etc.	$\sim 1.0 \times 10^{-9}$	Photochemistry	Photochemistry
HCl	$\sim 1.0 \times 10^{-9}$	Acidification	Rainout
NH_3	$(0.1\text{-}1) \times 10^{-9}$	Biology	Photooxidation
HNO_3	$(0.05\text{-}1) \times 10^{-9}$	Photochem (NO_2)	Rainout
COS	5×10^{-10}	Biology	Photodissociation
CH_3Cl	5×10^{-10}	Biology	Photooxidation
NO, NO_2	$(0.2\text{-}5) \times 10^{-10}$	Comb., biology	Photooxidation
$(CH_3)_2S$	$\sim 4 \times 10^{-10}$	Biology	Photooxidation
CF_2Cl_2	3.7×10^{-10}	Industry	Photodissociation
SO_2	$\sim 3 \times 10^{-10}$	Comb., photochem.	Photooxidation
$CFCl_3$	2.2×10^{-10}	Industry	Photodissociation
H_2S	$\sim 2 \times 10^{-10}$	Biology	Photooxidation

*Values quoted are for dry air.

missions that a light cloud haze might extend well below the main cloud deck. On the other hand, any acid rain could evaporate, due to the very high temperatures at lower altitudes, before reaching the surface.

Several of the gases listed in table 3-3 are cloud progenitors, particularly SO_2, H_2S, and CO_S. Volcanic eruptions or reactions of H_2O and CO_2 with volcanic surface rocks yield CO_S, H_2S, S_2, and SO_2. Various photochemical reactions and reactions with H_2O convert these species to concentrated H_2SO_4 or elemental sulfur particles in the clouds. The H_2SO_4 evaporates at and below the cloud base, producing SO_3, which can then either recondense or be reduced to SO_2. Reactions of SO_2 with Ca^{2+} in rocks provide a sink that must be balanced by the volcanic and surface sources.

Both the Soviet and United States spacecraft have suggested the presence of lightning on Venus. Although one Soviet observation has been optical, all of the other observations have been of low-frequency radio "static." On Earth, this static has been correlated with visible lightning bursts. It is not as clear-cut on Venus, and great controversy surrounds the observations and their interpretation. Lightning, if present, may be an important energy source for the production of new chemical compounds not possible through normal solar photochemical processes.

We conclude this first part of our discussion with a short synopsis of our knowledge of the surface of Venus, as this is the most likely platform for the existence of extant or extinct life. Also, as mentioned many times above, surface-atmosphere interactions and outgassing from the interior through the surface play important roles in the formation of the atmosphere and the chemical cycles controlling atmospheric species.

We begin with a review of the global topography. (The following review is based primarily on Pioneer Venus Orbiter Radar Mapper data. The Magellan data were not available when this chapter was written.) Variations in the radius of Venus range from 6049 km to 6062 km, a spread of 13 km. Since the mean radius (which we use as a reference on Venus in the absence of "sea level") is 6051 km, these extremes are –2 km to +11 km about the mean. Although the elevated terrain comprises a number of separated components, much less in number than on Earth, it is dominated by a massive equatorial region the size of South America. The total relief on Venus (13 km) is about two-thirds that on Earth (20 km).

Of the total surface, 60% lies within 500 m, and 20% lies within 125 m of the mean radius. The planetary polar ellipticity is nearly zero (upper bound of 4×10^{-5}). The surface of the planet may be divided into three provinces: upland rolling plains, making up 65% of the surface, lying between 6051 km and 6053 km; highlands, about 8% of the surface, between 6053 km and 6062 km; and lowlands, about 27%, between 6049 km and 6051 km. Numerous dark circular features in the rolling plains province may be lava-filled impact basins. A "granitic" composition for the rolling plains has been inferred; thus, this province may represent most of the planet's ancient crustal material.

Aphrodite Terra, centered at latitude 5° south between longitudes 80 and 190° east, and Ishtar Terra, centered between latitudes 60 and 75° north at longitude 0°, compose most of the highland province and in many ways resemble continents on the Earth. Their highest points stand 11.1 and 5.7 km above the mean planetary radius, respectively. Aphrodite appears to be highly disrupted tectonically, and degraded. Ishtar is made up of an uplifted plateau and great volcanic construct and is the site of the highest point, 11.1 km, Maxwell Montes (latitude 63.8° north, longitude 2.2° east), on Venus. Note that its elevation above the Venus datum is greater than the height of Mount Everest on Earth above sea level, which is 8.8 km. Gravity and altimetry data indicate that the highlands are compensated isostatically, probably as a result of crustal thickening or lateral variations in the crust and mantle, that is, by either passive or dynamic mechanisms.

The lowlands province of the planet includes several crudely circular low areas with low relief within the highlands. All lowland regions may be covered by younger basaltic lavas that have filled depressions where the crust is thinner. The lowest point on Venus is in a rift valley or trench named Diana Chasma, at latitude 14° south and longitude 156° east, where the elevation is 6049 km, or 2 km below datum. In comparison with terrestrial depths, this trench is deeper than the Dead Sea Rift but is less than one-fifth the maximum depth of the Mariana Trench, which is 11 km below sea level.

An integrated global pattern of subduction troughs or mid-basin ridges, indicative of active global plate tectonism, has not been identified. However, complex ridge-and-trough regions east of Ishtar Terra and in southern Aphrodite Terra, and a tectonically disrupted region between Beta Regio and Aphrodite Terra, may be the result of large-scale crustal motion. Beta Regio appears to consist of two giant irregular shield volcanoes, Theia Mons and Rhea Mons. Their relief profiles (both features reach elevations of more than 4.5 km above the datum) and the presence of a summit depression aligned on an axial trough suggest a basaltic compostion. This interpretation is supported by actual measurement of basaltic compositon of rocks directly east of Beta.

Despite the major bulk similarities between Venus and Earth, geologically interesting differences in atmospheric composition, atmospheric and lithospheric temperature, and possibly mantle composition suggest that the rock cycle on Venus is very different from the rock cycle on Earth. Exposed rocks on the surface of Venus have been sampled and appear to be similar to common igneous rocks on Earth. If differences of atmospheric pressure and temperature with altitude, and the probable wind transport of weathered regolith, are taken into consideration, then it is possible thermodynamically for the minerals in these common rocks to be decomposed by reaction with the atmosphere. Existing data are consistent with weathered igneous rocks, or compacted and partially cemented sedimentary rocks, or both.

Venus' Early History

We started this chapter on planetary exobiology with the assertion that extant life on the surface of Venus is out of the question. We have presented conclusive and dramatic evidence to support this assertion. This now leads us to the possibility of *extinct* life on Venus. Was there a period of time in Venus' 4.5 billion year history when the planet possessed those Earth-like characteristics (oceans and moderate surface temperatures) necessary for life to exist?

We begin with the most fundamental, and as yet unanswered question: Did Venus form with a large inventory of water, or did it form dry? The hypothesis for a dry origin for Venus is a prediction of the equilibrium condensation model for planetary formation. This model assumes that the bulk of Earth's (and Mars') water was incorporated into the planet in the form of hydrated minerals, such as tremolite [$Ca_2Mg_5Si_8O_{22}(OH)_2$] or serpentine [$Mg_3Si_2O_5(OH)_4$]. Formation of such minerals is predicted thermodynamically at the relatively low temperatures thought to prevail in the solar nebula beyond the orbit of proto-Earth, but would have been precluded in the warmer regions near the orbit of proto-Venus. This model presumes that the cooling time of the solar nebula was slow compared with the time for the planetesimals to form, so that the material that condensed at a given radial distance from the center of the nebula would have had a nearly uniform composition.

The equilibrium condensation model predictions have been seriously questioned in recent years, for two good reasons. First, the model presumes that the planets formed exclusively from material that condensed from the nebula in their immediate vicinity. It has been shown that radial mixing of planetesimals during the accretion process could have resulted in substantial exchange of material formed in different regions of the nebula. Indeed, this model predicts that the four innermost, terrestrial planets should all be composed of the same material. In reality, there are known compositional differences between these planets, so the actual degree of mixing was probably less than predicted.

Another approach to this question is to ask whether it is possible to build the terrestrial planets from a combination of known meteorite types. This, again, presumes some nebular mixing. The carbonaceous chondrites have been suggested by some researchers as the source of Earth's volatiles. These meteorites contain water (about 10%, by weight) in the form of hydrated minerals, along with substantial amounts of carbon in saturated organic compounds ('CH_2' for short). Oxidation of this organic carbon by ferric oxides contained in such planetesimals would have yielded carbon dioxide, ferrous oxide, and water:

$$CH_2 + 3\ Fe_3O_4 \rightarrow CO_2 + H_2O + 9\ FeO$$

The H_2O/C ratio after oxidation is about 4.5.

Other researchers have suggested that Earth's volatiles were obtained from ordinary chondrites. These meteorites are less highly oxidized and have much lower volatile contents than do carbonaceous chondrites. Their carbon exists mainly in amorphous, elemental form. Oxidation of this carbon would have yielded CO_2 and elemental iron, but no water:

$$C + 2\ FeO \rightarrow CO_2 + 2\ Fe$$

Ordinary chondrites do, however, contain H_2O (about 2%, by weight) in hydrated minerals. The H_2O/C ratio is about 2.

If the carbon in Venus' 95-bar, CO_2 atmosphere was derived from one of these sources, some 50 to 120 bars of H_2O, roughly one-fifth to one-half of a terrestrial ocean, would have entered at the same time. Thus, such material would have to have been completely excluded from the neighborhood of proto-Venus for Venus to have formed dry. The equilibrium condensation model circumvents this problem by suggesting that Venus' CO_2 was derived from metal carbides dissolved in an H_2O-deficient iron matrix.

The second problem with the equilibrium condensation model is that it does not consider the effects of comets. A large number of comets are thought to have been scattered into the inner solar system as a result of outer, giant planet orbit perturbations, where they could have collided with the recently formed terrestrial planets. The flux of comets during the first several hundred million years of solar system history may have been 10^4 to 10^5 times greater than today. It has been calculated that the H_2O in the Earth's oceans could have been derived entirely from H_2O-bearing comets, if they were responsible for 10% of the impacts recorded on the moon (meteorites and asteroids providing the other 90%). Studies of recent lunar impacts suggest that 10-50% are due to comets and the remainder are from asteroids. Of course, there is no reason why the ratio of comets to asteroids should have been the same early in solar system history; however, if Earth gained even a small fraction of its H_2O in the form of a late cometary veneer, then Venus should have received a comparable amount.

There is an important possible difference in timing between the late veneer, the cometary model, and the other models for water acquisition. If Earth's and Venus' H_2O were provided by comets, then it would probably have been supplied over a period of several 100 million years. (The heavy bombardment of the moon apparently continued until 3.8 billion years ago.) In contrast, H_2O obtained from inner solar system planetesimals would have been incorporated into the terrestrial planets within the first 100 million years. The time scale for H_2O loss for Venus, discussed later, is of the order of 100 million years or even less. A "late veneer" model for Venus might, therefore, have never had its full complement of H_2O present at any one time. This could make it less likely that any of this water condensed to form oceans. The atmosphere at the time, however, would still have been very wet compared to today.

Another argument that has been used in favor of a dry origin for Venus is that it would have been impossible to get rid of large amounts of water. The proposed mechanism for water loss involves photodissociation of water vapor in the upper atmosphere of Venus, followed by escape of hydrogen to space and loss of oxygen by chemical reactions at the surface. Conceptually, this scenario is logical; however, there are potential difficulties when one examines the process in detail.

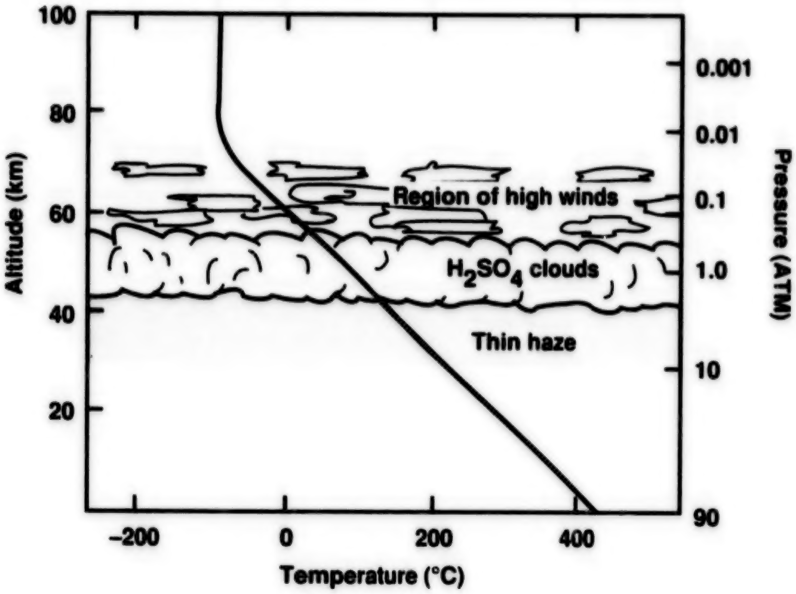


Figure 3-3. Venus' vertical temperature profile showing an atmospheric cold trap.

The suggested escape mechanism for hydrogen during the early stages of water loss involves hydrodynamic outflow. Theory indicates that the hydrodynamic outflow is highly efficient if the upper atmosphere was hydrogen-rich. However, there is a catch—water vapor, hence hydrogen, could have been effectively confined to the lower atmosphere by an atmospheric cold trap (fig. 3-3). The cold trap is that region of the atmosphere where the fractional concentration of water vapor is held to a minimum by condensation. In the Earth's atmosphere the cold trap occurs between 9 and 17 km altitude; it is coldest and thus most effective near the equator. Earth's cold trap limits the concentration of water vapor in the stratosphere to only a few parts per million by volume. The escape rate of hydrogen from Earth's atmosphere is consequently far too low to affect the amount of water stored in the oceans.

Early Venus would have been different. Climate models predict that a cold trap does not work well when the lower atmosphere contains more than about 10% water vapor by mass. A wet early Venus would have at least this much water vapor in its atmosphere. When so much moisture is present, the amount of latent heat released by condensation and cloud formation is so large that the cold trap moves up to very high altitudes. There, the ambient pressure is comparable to the saturation vapor pressure of water, so condensation has little effect on the water concentration. Water vapor can, thus, make its way unimpeded into the upper atmosphere, where it can be photodissociated and the hydrogen lost to space.

J. S. Lewis has focused on the difficulties of disposing of the oxygen left behind after the hydrogen has escaped, to support his dry-origin Venus theory. In the carbonaceous chondrite model for terrestrial planet formation, Venus would have been left with approximately 110 bars of O_2 from its initial 120 bars of H_2O. If this oxygen was consumed in oxidizing ferrous oxide to magnetite:

$$3\ FeO + 1/2\ O_2 \rightarrow Fe_3O_4$$

and if Venus' crust was approximately 10% FeO (like Earth), some 80 km of crustal rock would have been needed to take it up. This would require exposing 60 km^3 of fresh material each year, an amount 15 times greater than the volume of new crust created annually at the midocean ridges on Earth by plate tectonics.

Whether this poses a problem for the wet origin model depends in part on when the water was acquired. If much of the water came in during the accretion period itself, then the planet's surface would have been molten and the entire mantle should have been convecting vigorously. The amount of oxygen that could have gone into the mantle under these circumstances is virtually unlimited.

Indeed, water would probably have reacted with elemental iron in the melt and generated hydrogen directly:

$$H_2O + Fe \rightarrow H_2 + FeO$$

This hydrogen would have been outgassed and made its way to the top of the atmosphere unhindered by condensation. If sufficient solar extreme ultraviolet (EUV) energy was available to allow it to escape, it would have done so at that time. If the inner solar system was still filled with dust from planetesimal collisions obscuring the EUV light, the hydrogen would have remained in Venus' atmosphere until the nebula cleared and then escaped. In either case, large amounts of water could have been lost without creating any free oxygen.

Disposing of cometary water acquired after the main accretion period would have presented a bigger problem. Roughly 1/30 of a terrestrial ocean (or an average depth of 100 meters) could have been disposed of by oxidizing an amount of fresh crustal material comparable to that presently generated on Earth. This figure could be multiplied by a factor of two or three if Venus, like Earth, was more tectonically active in the past.

But there are other possible sinks for oxygen that could accommodate large quantities of cometary water. For example, Venus' CO_2 may have been originally outgassed as CO instead of being released in a fully oxidized state as the reactions above would suggest. Some 30 bars of H_2O, or one-tenth of a terrestrial ocean, could have been consumed in oxidizing this CO.

A second possibility is the escape of oxygen to space. A hydrodynamic hydrogen escape flux in excess of 2×10^{13} H atoms cm^{-2} s^{-1} would have been sufficiently vigorous to drag some oxygen atoms along with it. Such an escape rate is energetically possible on Venus during the first 500 million years of solar system history, given an enhanced solar EUV flux at that time. (Large EUV enhancements have been predicted for the young Sun based on observations of T-Tauri stars and on stellar evolution theory.) Indeed, from an energetic standpoint it is possible for Venus to have lost several oceans of water, including the oxygen, during the first 100 million years of solar system history. The actual efficiency would depend on the opacity of the inner solar sytem to EUV during this time period.

The real Achilles' heel of the Runaway Greenhouse Effect hypothesis lies in getting rid of the last part of the original water endowment. Venus' water would have been lost readily until the mass mixing ratio of water vapor in the lower atmosphere had fallen below 0.1. At this point, a cold trap should have developed, blocking water vapor transport to the upper atmosphere. If Venus had its massive 95-bar CO_2 atmosphere at this time, roughly 10 bars of H_2O would have remained in its lower atmosphere. It is difficult to estimate exactly how fast this water could have escaped, but it may be impossible to lose this much water even over the course of several billion years.

The problem is compounded further by sulfur photochemistry. At some stage in the water loss process, sulfuric acid clouds would have started to form. These clouds are extremely hygroscopic (water-absorbing) and could have dried out the Venus upper atmosphere even more, further reducing the escape rate.

Thus, the classical runaway greenhouse model for Venus, in which the water is all in the vapor phase, encounters significant problems in losing the water. This has led one of us (Kasting) and his colleagues at NASA's Ames Research Center to revisit the climate models to try to resolve this problem.

If one takes an Earth-like planet with an ocean-covered surface and calculates how much solar heating is required to completely vaporize that ocean, Kasting's climate model predicts that you need a flux about 1.4 times greater than the current flux (S_0) at Earth's orbit. This calculation assumes no change in cloud cover. If cloudiness increases with increasing surface temperature, as seems likely, the critical solar flux could be considerably higher. The current flux at Venus' orbit is 1.91 S_0, well above Runaway Greenhouse limit of 1.4 S_0. However, the Sun was approximately 30% less bright shortly after it was formed, so the flux incident on primitive Venus was only 1.34 S_0. This is close enough to the Runaway Greenhouse limit to lie within the uncertainty of the model calculation. However, the inclusion of cloud feedback would cool the planet considerably and make a true Runaway Greenhouse Effect unlikely.

Hence, if Venus did start out with an Earth-like water endowment, much of that water should have condensed to form a hot ocean. The temperature of that ocean depends on the climatic effect of the clouds and on the amount of CO_2 present, but it would likely have been between 100 and 200°C. The corresponding vapor pressure of water is 1 to 15 bars. A fully vaporized terrestrial ocean, by comparison, would produce a surface pressure on Venus of about 250 bars. Liquid water should therefore have been stable on early Venus even if the total water endowment was only a fraction of Earth's. We call this modified, ocean-stable greenhouse model the "Moist Greenhouse" to distinguish it from the oceanless Runaway Greenhouse.

There is more than just semantics or labels involved here. The Moist Greenhouse model leads to very different predictions concerning Venus' early history. The presence of an ocean on early Venus should have caused large changes in the composition of the atmosphere. On Earth, water provides a medium for weathering silicate materials and converting them into carbonates.

Atmospheric CO_2 is consumed in the process. The simplest such reaction involves wollastonite reacting with the carbon dioxide to produce calcite and quartz:

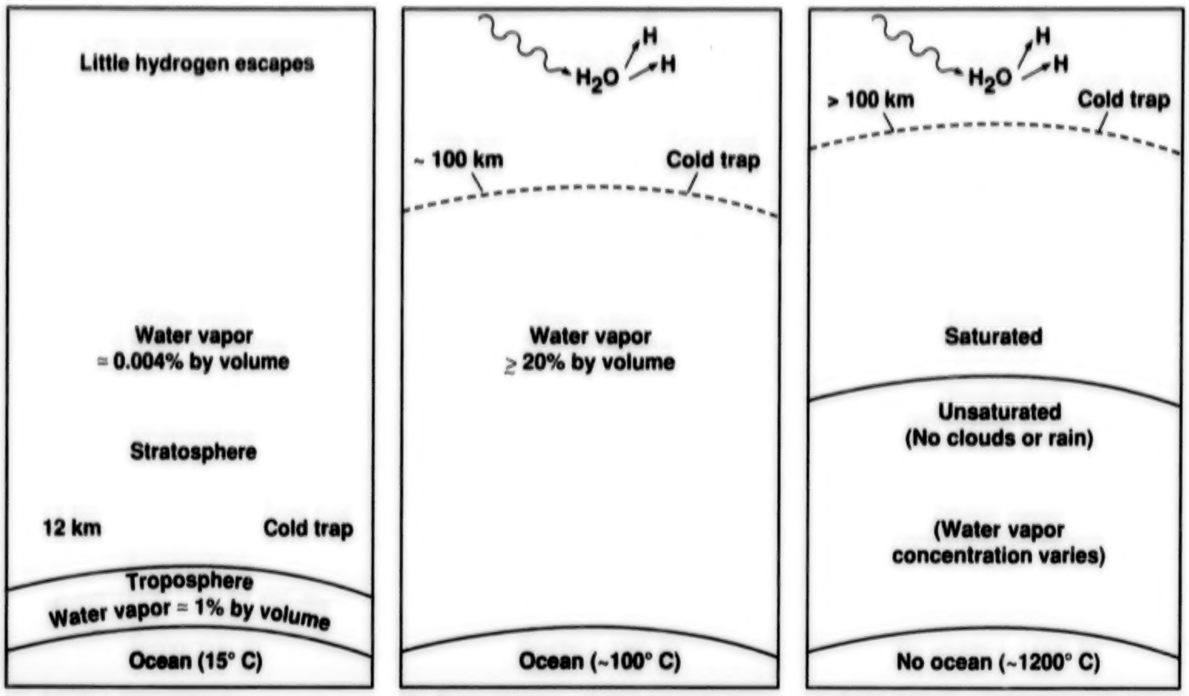
$$CaSiO_3 + CO_2 \rightarrow CaCO_3 + SiO_2$$

Reactions like this, which occur readily in the presence of liquid water, would have reduced the atmospheric pressure by sequestering CO_2 in the planet's crust.

Somewhat counter-intuitively, this reduction in atmospheric CO_2 should have facilitated the escape of water. Suppose, for example, all of Venus' CO_2 was converted into carbonates in this manner. The remaining atmosphere would have been a mix of roughly 2 bars of N_2 plus however much water vapor was present at saturation. If the surface temperature was 100°C or higher, the concentration of water vapor in the lower atmosphere would have been 25% by mass, or more. The cold trap would have been ineffective, and hydrogen would have escaped from the top of the atmosphere at a rapid, hydrodynamically controlled rate. This rapid water loss would have continued until the water concentration dropped below about 10% by mass, at which point only about 0.2 bar of water would have remained in the atmosphere—the ocean should have already evaporated by this time. Because this atmosphere is 50 times thinner than the present one, some 50 times less water would remain after the cold trap formed and the hydrodynamic escape process stopped.

The presence of liquid water would also have helped to solve the problem of the water-trapping sulfuric acid clouds. All of the common sulfur gases—SO_2, H_2S, H_2SO_4—are soluble in water to some extent. If an ocean were present, they would have dissolved to form sulfite, sulfide, and sulfate. These species, in turn, would have combined with available cations to form various sulfur-containing minerals. The sulfuric acid clouds could not have formed until the ocean had dissappeared and sulfur was recycled into the atmosphere by volcanic activity. CO_2 would have been regenerated in a similar manner. Since Venus, like Earth, was producing heat in its interior, its mantle must have been convecting and its surface must have been reprocessed by some form of tectonic activity: point volcanism, perhaps, if not plate tectonics. If such reprocessing was occurring, carbonate rocks would have undergone metamorphism and gaseous CO_2 would have been recycled back into the atmosphere. Over billions of years, volcanic outgassing of CO_2 and SO_2 would have caused the atmosphere to evolve to its current state.

Let us recap (fig. 3-4), then, a reasonable theory for the history of water on Venus. Venus started off wet because it would have received a certain percentage of the same volatile-rich material which formed the Earth. Once the initial accretion period was over, the combination of lower solar luminosity and a high albedo caused by the clouds would have resulted in a relatively cool surface temperature. If Venus had anything approaching Earth's water inventory, much of this water would have condensed to form oceans. Carbon dioxide that was originally present in its atmosphere would have been slowly converted to carbonate rocks, and the atmosphere would have thinned. Water would


Little hydrogen escapes
Water vapor
≈ 0.004% by volume
Stratosphere
12 km
Cold trap
Troposphere
Water vapor ≈ 1% by volume
Ocean (15° C)
H
H₂O
H
~ 100 km
Cold trap
Water vapor
≥ 20% by volume
Ocean (~100° C)
H
H₂O
H
> 100 km
Cold trap
Saturated
Unsaturated
(No clouds or rain)
(Water vapor
concentration varies)
No ocean (~1200° C)

(a) Earth's atmosphere **(b) Early Venus – moist greenhouse** **(c) Early Venus – runaway greenhouse**

Figure 3-4. Tendency of water vapor to escape from the Earth is minimal; the same cannot be said for early Venus. On the Earth (a) water in the troposphere is blocked from entering the stratosphere by a cold trap, the region where cold temperature and relatively high ambient pressure combine to minimize the concentration of water vapor. When vapor reaches the trap, most of it condenses out. On early Venus the lower atmosphere, though warm by the Earth's standards, may have been cool enough for water to condense and form an ocean. The sea would in time have been lost, however, to a "Moist Greenhouse" (b), a condition that arises when a high surface temperature enables water vapor to constitute more than about 20% of the lower atmosphere. The cold trap then moves to a high altitude and becomes inefficient at preventing water vapor from rising into the upper atmosphere. Although some vapor condenses out as rain, the steam at the top dissociates and its constituent hydrogen atoms escape into space. Venus might have been so hot that a Runaway Greenhouse (c) developed instead; all the water released by the planet turned to steam instantly, and no ocean formed. The water essentially traversed a one-way route: up and away.

have remained a major atmospheric constituent throughout this period, its abundance gradually decreasing with time as a consequence of photodissociation followed by hydrogen escape. Some of the oxygen released by this process may have been dragged off into space along with the hydrogen; the rest would have been consumed by oxidizing carbon monoxide and by reactions with reduced minerals (primarily ferrous oxide) in the planet's crust. Because the atmosphere was thinner than it is today, hydrodynamic escape would have removed all but a few tenths of a bar of Venus' original water endowment.

The remainder was lost over billions of years by slower, non-thermal escape processes. The disappearance of water allowed the CO_2 and SO_2 released by volcanoes to accumulate, and the atmosphere gradually approached its present state.

So it appears to be quite possible that Venus possessed a sizable water inventory and a cooler climate (100-200°C) in its early history, perhaps for as long as a few hundred million years. Even 100°C would probably still be considered too hot for life to originate from the viewpoint of many exobiologists. On the other hand, Earth may have been much hotter than today when life formed on this planet. A more serious problem may have been the effects of late impacts. If the impact rate for Venus was as high as we think it was for the Earth and the Moon, life may have been repeatedly wiped out even though other environmental conditions were favorable.

Is there experimental evidence for the existence of water in large quantity on Venus during its past? There is a positive result inferred from the mass spectrometer experiment on-board the Pioneer Venus Large Probe. As the probe descended through the clouds of Venus on December 9, 1978, its inlet became clogged, apparently by a large H_2SO_4 cloud particle, for a period of time. During the time the instrument was "failing," two groups of experimenters were able to analyze in detail the cloud particle trapped in the spectrometer. They were able to deduce the deuterium (heavy hydrogen) to normal hydrogen ratio, and found it to be 1.0 to 1.6×10^{-2}, or about 100 times higher than the same ratio on Earth.

They concluded that the present ratio on Venus is a residue from selective escape of at least 100 times the current water abundance. In other words, if the original D/H ratios were the same for Earth and Venus, the current high ratio on Venus is due to the fact that the lighter, normal hydrogen escaped more readily than the heavier deuterium. Of course, 100 times the present water abundance would still be only about 0.1% of a full terrestrial ocean. However, Earth-sized oceans on Venus are possible if there was significant deuterium escape as well.

These inferences have recently been questioned. The formation of the inner planets is not understood well enough to assume that the original ratios for Venus and Earth were about the same. Furthermore, the high ratio may not be evidence for Venus' proposed oceans anyway. The water seen today could have resulted from the sporadic infall of cometary material for which the D/H ratio is largely unknown; the ratio was determined for comet Halley and is terrestrial within a factor of three. These calculations assumed a water abundance of 10 parts per million; they do not work if the abundance is above 200 parts per million.

Thus, this indirect evidence for early Venus oceans is highly controversial. More direct evidence, e.g., evidence of fluvial channels as seen on Mars in Viking orbiter images, has not been reported on Venus.

What kind of experiments should be designed to look for evidence of extinct life on Venus? Probably they should be similar to those being planned to look for evidence of extinct life on Mars on the proposed Mars Rover Sample Return (MRSR) mission. The technology for a MRSR on the surface of Mars, with its thin, cold atmosphere, is available today. The technology for a Venus Rover Sample Return mission on the the surface of Venus is beyond our capabilities and will probably remain so for two or three decades, at least.

Prior to the advent of the space age, Venus was often referred to as Earth's "twin." This was the result of the similarities in bulk properties (size, mass, density) and the fact that Venus is our nearest planetary neighbor. As spacecraft revealed the remarkable differences between the two planets, the appelation lost its relevance. However, the recent studies of early Venus and early Earth have revealed potentially new similarities, and the new tectonic inferences add to this picture. Thus, we may be forced to revive the twin analogy—perhaps twins at birth that evolved along substantially different paths to maturity.

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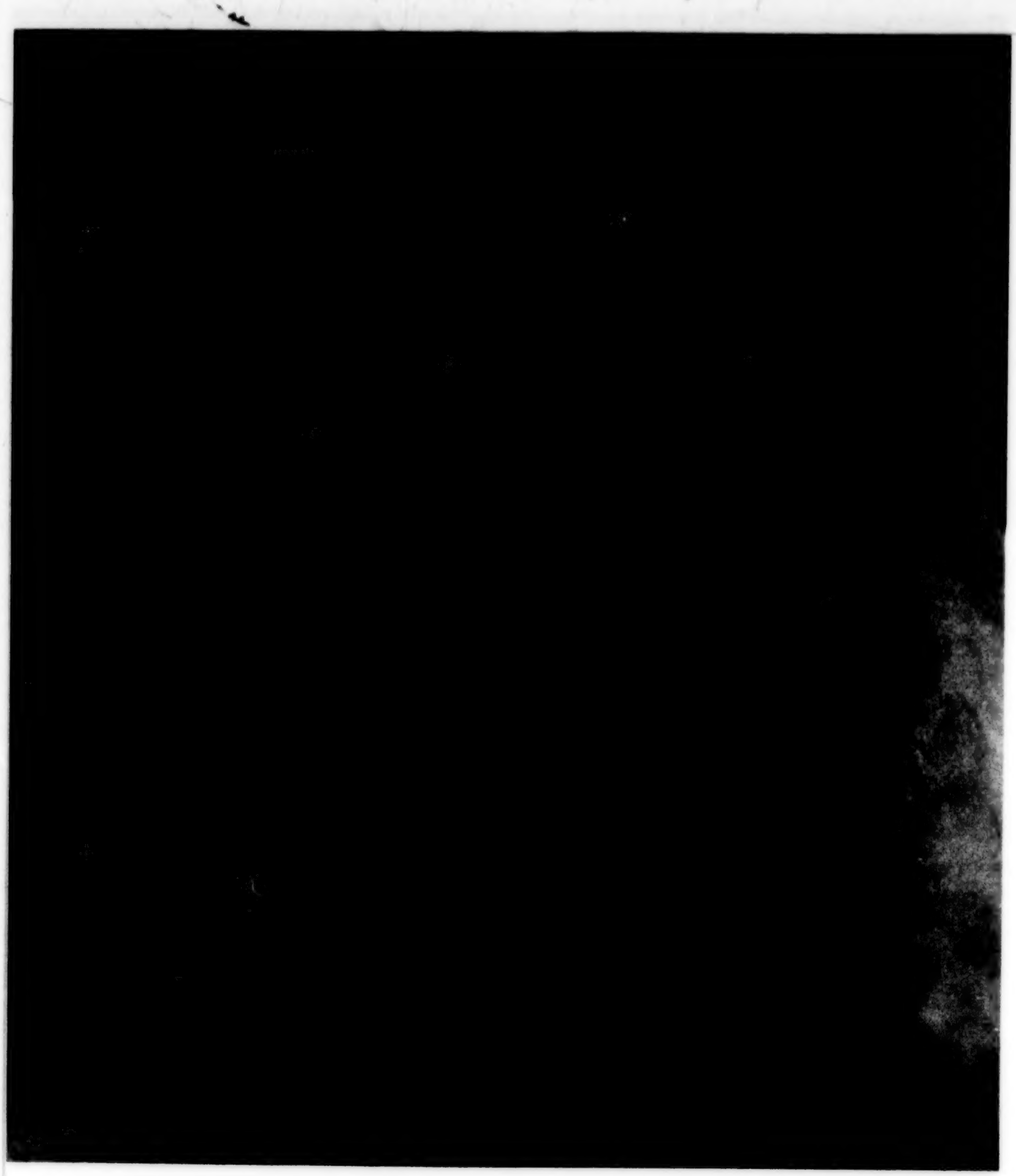

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Mars: A Reassessment of its Interest to Biology

C. P. McKay

Of all the other planets in our solar system, Mars is certainly the one that has inspired the most speculation concerning extraterrestrial life. Looking through telescopes, observers had long noticed that Mars exhibits changes in its polar caps and alterations in its surface coloration that parallel seasonal changes on Earth. Seasons result from the axial obliquity, and the Martian obliquity (25°) is very close to that for the Earth (23.5°). Mars is the only planet in the night sky which exhibits such an Earth-like pattern of seasonal change. This similarity may have been the basis for the early speculation that Mars was the home to civilizations: ancient, warlike, wise, or on the eve of destruction—depending on the author's inclination.

The fascination with Mars and the possibility of life on Mars continued into the spacecraft era and was directly expressed in the Viking Missions. These highly successful missions had the search for life on Mars as one of their principal goals. To conduct the search two identical

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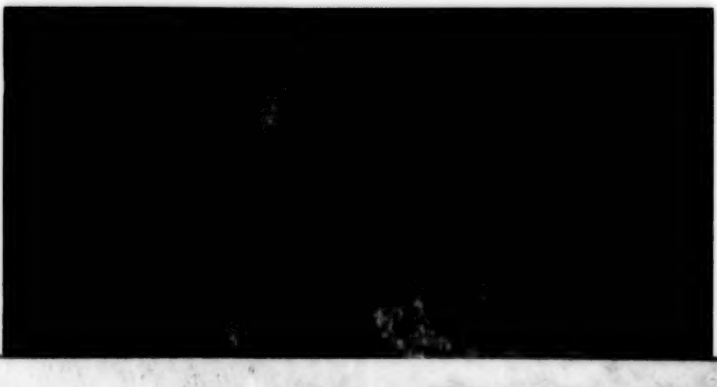


landers set down on opposite sides of the Martian northern hemisphere.

In addition to the lander cameras, which would show the presence of any obvious macroscopic life-forms, and the GCMS (Gas Chromatograph/Mass Spectrometer) which searched for organics in the soil, the Viking landers contained three experiments specifically designed to search for indications of life on Mars: The Gas Exchange (GEX) experiment was designed to determine if Martian life could metabolize and exchange gaseous products in the presence of water vapor and in a nutrient solution. The Labeled Release (LR) experiment sought to detect life by the release of radioactively labeled carbon initially incorporated into organic compounds in a nutrient solution. The Pyrolytic Release (PR) experiment was based on the assumption that Martian life would have the capability to incorporate radioactively labeled carbon dioxide in the presence of sunlight (i.e., photosynthesis). In addition, the X-ray Fluorescence experiment analyzed the elemental composition of the loose material at the Viking lander sites. Unfortunately, from the biology perspective, the instrument could only detect elements with atomic number greater than that of Mg. Thus there was no direct measurement of O, N, C, or H in the soil material.

The Viking results indicated that the surface of Mars is a cold dry desert. The thin atmosphere (about 8 mb surface pressure) is composed of 95% CO_2 with 1.5% N_2 present. Water is also present in the atmosphere at about 0.03% and in the soils at the Viking lander sites with concentration of a few percent by mass. Phosphorus was not detected because its signal in the X-ray Fluorescence instrument was hidden by that of S and Si—both of which were present—but is nevertheless thought to be present. Thus all the biogenic elements (C, H, N, O, P, S) are present on the surface of Mars.

Perhaps the most surprising result of the Viking soil analysis was the virtual lack of organic carbon in the soil. The GCMS failed to detect organics in surface samples and from samples below the surface (maximum depth sampled was about 10 cm) at parts per billion levels. There are at least two mechanisms that could produce organics on Mars. One mechanism is the importation of organic material by the infall of meteorites, many of which are known to carry organic material. A second mechanism is the production of organic material by UV light incident on the surface, as demonstrated in the Viking Pyrolytic Release Experiment. Hence, the absence of organics suggests that a mechanism for destroying them is present. All that the GCMS did detect were traces of the cleaning fluids used in preparing the instrument before launch. The lack of organics in the Martian soil is certainly a persuasive argument against the presence of life at the Viking landing sites.

Yet, the Viking Biology instruments gave interesting results. In the GEX experiment, the soil released O_2 upon humidification in amounts ranging from 70-790 nmoles cm^{-3}. Heating of the sample to 145°C for 3.5 hours reduced the amount of O_2 released by about 50%. There was a slow evolution of CO_2 when nutrient was added to the soil. In the

LR experiment there was a rapid release of CO_2, followed by a prolonged slow release of CO_2, from radioactively labeled C in the nutrient solution. The effect was completely removed by heating to 160°C for 3 hours, partially destroyed at 40–60°C, and relatively stable for short periods at 18°C, but was lost after long-term storage at 18°C.

The chemical activity and lack of organics is likely caused by reactions with one or more oxidants in the Martian soil. The chemical composition of these oxidants is not certain. Based upon the GEX and LR results, there must be at least three oxidants on Mars. The GEX results imply the existence of a strong oxidant that is thermally stable and capable of breaking down water to release oxygen. A second strong oxidant must exist to explain the LR results. This oxidant differs from the GEX oxidant in that it is thermally labile. A third, weak, oxidant (γ-Fe_2O_3) is required to explain the slow oxidation of the nutrient in the GEX experiment and the release of CO_2.

Possible oxidants for the GEX oxidant include KO_2, ZnO_2, and CaO_2. These oxidants would need to be present at concentrations of about 2-25 ppm by mass to match the experimental results. A possible oxidant for the LR oxidant is H_2O_2, catalyzed in the surface by the soil minerals. H_2O_2 could be produced in the atmosphere from photochemical reactions at a rate of about 2×10^9 molecules cm^{-2} s^{-1} and could be the source of the oxidants in the LR experiment. The concentration of H_2O_2 required to explain the LR results is about 1 ppm by mass.

Alternative explanations for the Viking biology experiments include the suggestion that the chemical reactions were due to peroxonitrite or to intrinsically reactive clays or the production of radicals such as OH^- resulting from chemical weathering processes in the soil without any UV excitation. It has been suggested that the O_2 released in the GEX experiment could be due to physically trapped O_2 within micropores or the release of physically absorbed gases in the Martian soil, with no chemical reactions necessary to explain the results.

Thus, the Viking Missions returned a negative, but still equivocal, answer to the question of life on Mars. Extending the unfavorable results to the rest of the Martian surface may not be justified given our present state of knowledge of Mars. Furthermore, the two Viking sites were similar in that they were both covered by a mantle of windblown dust. Although this dust layer appears to be ubiquitous on Mars, localized spots may exist that have quite different conditions than the Viking lander sites. However, in at least two respects the prospects for life anywhere on Mars is diminished. First, if the oxidants that were thought to have been detected at the Viking lander sites are produced by photochemical reactions then they should be found everywhere on the planet with the concomitant destruction of organic material. Second, but more seriously, the amount of water anywhere on Mars is quite low and there is no indication that liquid water is found anywhere at any season. Without liquid water there is no prospect for the growth of life as we know it; for example, bacteria require water activities above 0.6 for metabolism.

While the spacecraft exploration of Mars has diminished hopes of finding extant life forms on its surface, it has also shown us that many years ago Mars was a very different place and enjoyed conditions that may have been conducive to the origin of life—life that may have long since become extinct.

From a biological perspective, the most important information returned from spacecraft exploration of Mars may be the geological evidence that liquid water was abundant on the Martian surface at some time in the past. The most compelling evidence for liquid water comes from the observation of riverine features. These include the outflow channels and valley networks whose morphology is characteristic of formation by liquid water. Further evidence that Mars is rich in water includes patterned ground and topographical features which suggest the fluid flow of soil material indicating near-surface ice. The presence of these fluvial features is a puzzle because liquid water is unstable on the present Mars primarily because the total atmospheric pressure is close to (and sometimes below) the triple point pressure of H_2O.

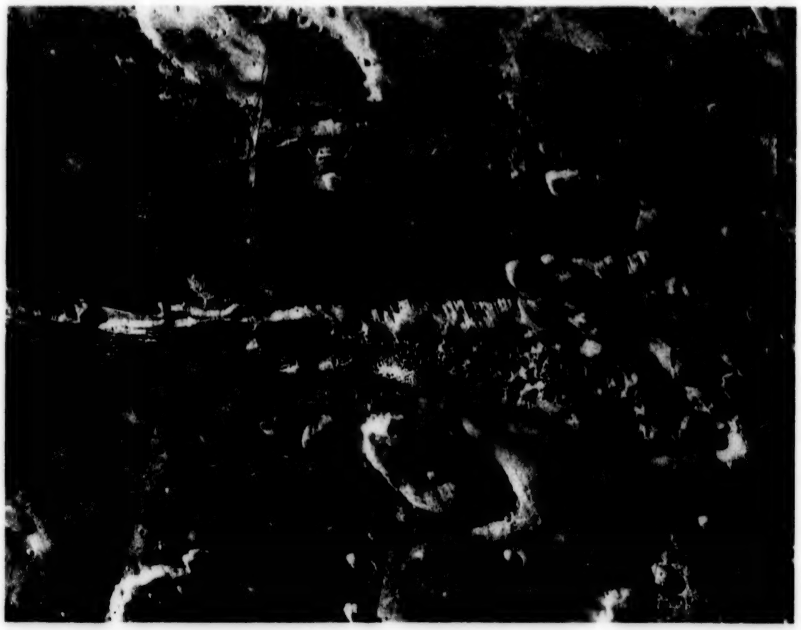


Figure 4-1. An outflow channel (Ravi Vallis) located at 1°S 42°W. The channel is 20 km across and appears to originate full-born from the enclosed region of chaotic terrain. These channels are thought to have been formed by the rapid release and flow of large quantities of water, thereby suggesting that Mars does have a significant inventory of water. Such channels could form under current climatic conditions.

The outflow channels and the valley networks provide two complimentary pieces of information about water on Mars. As illustrated by the size of the outflow channel in figure 4-1, these features are indicative of large-scale fluvial processes that must have been caused by catastrophic flooding events. They may have been caused by the rapid drainage of ice-dammed underground reservoirs. The rush of water under these conditions is such that these features could form even under the current Martian climatic conditions. To form these systems, large amounts of liquid water must have flowed on the Martian surface. The fluvial erosion indicated in the outflow channels implies a layer of water on Mars from 0.5 to 1 km thick.

On the other hand, the valley networks (fig. 4-2) appear to have been caused by gradual erosion due to slowly running water. Many of these channels could have been caused by the melting of ground ice or the release of groundwater. However, the dendritic drainage systems often associated with the valley networks could indicate rainfall. From the size and length of some of these networks it is clear that water must have been fairly stable at the surface. The valley networks are commonly found in the ancient cratered terrain in the southern hemisphere, generally thought to be the oldest Martian terrain, and are rarely found on the younger northern plains. This would suggest that the networks are old; they are believed to have formed primarily during, and shortly after, the decline in impact rates some 3.8 billion years ago. However, there is some fragmentary evidence to suggest that small, water-carved channels have been formed periodically throughout Martian history. The source of the water for these drainage systems is not clear: rainfall, glacial melt, and groundwater have been suggested.

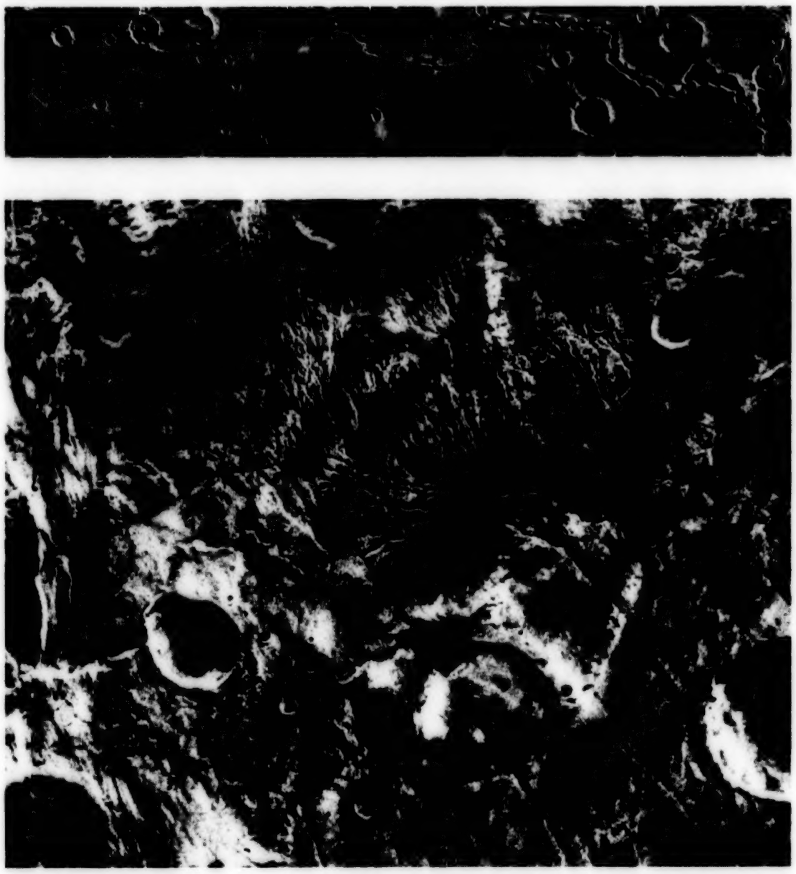


Figure 4-2. Runoff and dendritic channel. The runoff channel, Nirgal Vallis (top panel), located at 28°S, 40°W, is 800 km long and looks different from rivers on Earth because of the open nature of the network and the lack of a large catchment area; such channels probably derive from groundwater sapping rather than surface runoff. Dendritic channels (bottom panel) found in the ancient terrain are most probably formed by surface runoff following precipitation. The craters overlying the channels indicates that these features formed about 3.8 billion years ago. Unlike the outflow channels (fig. 4-1), the runoff and dendritic channels were probably formed under a significantly warmer climate—presumably caused by a thicker atmosphere—than the present Mars. This evidence for the stability of liquid water in the Martian surface 3.8 billion years ago is the primary motivation for considering the possible origin of life on Mars.

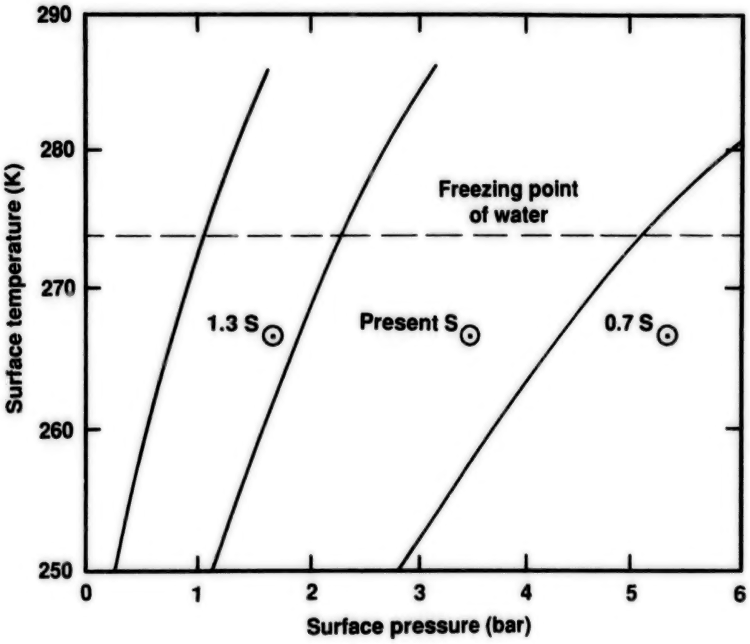
The sheer volume of water implied by the presence of outflow channels suggests that there was a significant amount of water on Mars and the long complex flow patterns of the valley networks indicate that, at some time, this liquid water was quite stable on the surface. While the amount of water present is unclear, it seems certain that there must have been aquatic habitats on early Mars. Water is the quintessence of life and the determination that there was liquid water on the surface of Mars for an extended period in its early history is the pivotal point upon which the discussion of a possible past Martian biota rests.

In order for liquid water to have freely existed on the surface of Mars in the past, conditions must have been quite different from the present. For liquid water to be stable, a thicker atmosphere would have been required with mean surface temperatures at or above freezing. Carbon dioxide is considered to have been the major constituent of the early Martian atmosphere, as it is in the present atmospheres of Venus and Mars. Similarly, it is believed to have been a major constituent in the early atmosphere of the Earth. Increased CO_2 in the atmosphere of early Mars could have provided the greenhouse effect required to keep the temperature above freezing even though the early Sun was 30% dimmer than at present.

A one-dimensional radiative-convective model of the early Martian atmosphere has been used to determine the amount of CO_2 required to raise the surface temperature above freezing, under the conditions of the faint early Sun. The calculations were based on an atmosphere composed of pure CO_2 in equilibrium with liquid water. The results, shown in figure 4-3, indicate that 5 bars of CO_2 were required to raise the globally and seasonally averaged surface temperature above the freezing point of water. At times of maximum eccentricity, one bar is required to raise the temperature above freezing at the equator at perihelion, producing seasonal meltwater.

It is possible that the mean temperature was below freezing and that only at certain places and during

Figure 4-3. Surface temperature as a function of CO_2 surface pressure. Globally averaged results are shown for the present level of sunlight and for faint early sun (30% reduction). Also shown is the result for the subsolar point on early Mars. These results suggest that about 1-5 bars of atmospheric CO_2 was required to raise the Martian surface temperature above freezing, as indicated by the fluvial features shown in figure 4-2.



certain seasons did temperatures rise above freezing. Valley networks could be formed while the mean temperature was many tens of degrees below freezing during seasonal warm periods. This suggestion is based on studies of the dry valleys of the Antarctic. The annual mean temperature in these valleys is –20°C, but the few days each year that the temperature is above freezing is adequate to form stream beds and ice-covered lakes. To have warmed only the perihelion subsolar point on the early Mars above freezing, presumably the warmest point and time on the planet, only 0.75 bars of CO_2 would have been required. Peak temperatures above freezing could have been maintained about 700 million years after the mean annual temperatures fell below freezing. Ice-covered lakes, such as those found in Antarctica, could have persisted during this interval.

If a thick CO_2 atmosphere was present on Mars, then it is reasonable to expect that there was a corresponding level of N_2. Estimates for the total amount of N_2 on early Mars range from 2 to 300 mb and, thus, cross the lower limit of N_2 pressure at which terrestrial organisms can incorporate N_2 by nitrogen fixation. The current partial pressure of N_2 of 0.2 mb appears to be well below the limit. Thus nitrogen, an essential macronutrient, was probably available on early Mars, but could have been a seriously limiting nutrient at later stages in the evolution of the atmosphere.

In addition to liquid water and a thicker atmosphere, Mars had much increased volcanism early in its history. The presence of volcanic events on early Mars may have played a key role in maintaining the environment as well as providing a habitat for early organisms. Mars, at present, is seismically and volcanically inactive when compared to the Earth, although there is ample evidence for tectonic and volcanic activity in the past. Most notable are the Tharsis Bulge and the large shield volcanoes, such as Olympus Mons (27 km above the mean elevation), which are found atop Tharsis. There is some evidence that volcanic activity has occurred, albeit at a reduced level, for most of Martian history. There is even the intriguing possibility that volcanism may have occurred in the Valles Marineris in geologically recent times.

The picture of the early climatic conditions on Mars that emerges from the previous discussion is one that is similar to early Earth: Both planets had liquid water on the surface, relatively thick CO_2–N_2 atmospheres, and volcanic activity. The presence of life early in Earth's history and the similarity between early Earth and early Mars thus motivates the question of early life on Mars.

As mentioned above, the conditions on the early Earth and early Mars were similar in many of the ecologically important parameters. A logical approach, then, to the question of possible past life on Mars is to investigate the history of early life on Earth. By looking at the record of early life on Earth we can develop concepts that may be applicable to life on early Mars and develop the tools to search the Martian sediments for fossil evidence of past life.

On Earth, life appears to have become widespread and to have developed sophisticated ecological communities by 3.5 billion years ago. Oxygenic photosynthesis by cyanobacteria may have even existed at this time. However, there is not general agreement as to when oxygenic photosynthesis developed. Pushing the origin of life further back into the geological past is difficult due to the scarcity of unaltered rocks of age 3.5 billion years and older. One intriguing set of sediments is found in Isua (Greenland) which dates back to 3.8 billion years ago. The organic material recovered from these sediments is also consistent with biological activity suggesting an even earlier origin for life. Thus, the time interval for the origin of life on Earth is between approximately 4.2 (the time of solidification of the mantle) and 3.5 billion years ago and life may have evolved over a much shorter time period.

To argue for the origin of life on Mars by analogy with the origin of life on Earth, the critical unknown is how long clement conditions prevailed after the first occurrence of liquid water on the Martian surface compared to the time required for life to have originated on Earth.

It has been suggested that the early thick Martian atmosphere was short lived. The fluvial features, particularly the valley networks, are evidence of a thick, moist early atmosphere. However, in this view, the atmosphere could not have been very thick for any significant period of time after 3.8 billion years ago. This argument is based upon the low levels of erosion, and the absence of infilling, of old post-bombardment surfaces and craters. Although erosion rates may have been larger before and just after the end of the early bombardment, they must have declined sharply with time. This is an important geomorphological argument *against* a thick, early Martian atmosphere lasting for many billions of years.

Mars would have lost its atmosphere as the CO_2 was transformed into carbonate rocks or was absorbed into the regolith. The timescale for eliminating atmospheric CO_2 by carbonate formation on early Mars is estimated to be a few times 10^7 yr. Thus, in the absence of recycling, the lifetime of a thick early atmosphere would have been very short indeed. On an active planet like the Earth, subduction of ocean sediments at plate boundaries followed by decomposition of carbonates in the mantle is the primary mechanism for completing the long-term geochemical CO_2 cycle. Mars does not have sufficient heat flow at present to cause the global scale recycling of volatiles incorporated into crustal rocks, nor is there any sign that Mars has, or ever had crustal dynamics akin to plate tectonics—rather its features are consistent with a 1-plate planet. Without these processes there appears to be no long-term geological mechanism on Mars to recycle CO_2-sink materials back into the atmosphere.

Perhaps intensive volcanism, driven by the high heat flows on early Mars, would have buried carbonate rocks to depths corresponding to their decomposition temperature, causing subsequent outgassing of CO_2, allowing for partial recycling. Eventually, as the interior of Mars cooled, the rate of volcanism would have been unable to recycle carbonates as fast as they were created, and the atmospheric pressure would have dropped. With this scenario, a thick moist CO_2 atmosphere could be retained for up to 10^9 yr, depending upon estimated values of the primordial heat flow and the total CO_2 budget on early Mars. Decomposition by impacts may have been a more effective mechanism for recycling carbonate rocks on early Mars. This mechanism would tie the existence of the thick CO_2 atmosphere to the impact history and would be consistent with the previously mentioned decline in erosion rate (and atmospheric pressure) after the end of the early bombardment.

It is also possible that Mars lost a considerable fraction of its early atmosphere as a result of erosion by high velocity impacts.

It was recently suggested that the flanks of Alba Patera contain channel features that are clearly of fluvial origin, and that the time of formation of these channels was well after the termination of the late bombardment on Mars. This implies these channels could not have formed as part of the putative warm, moist early epoch—which they presume to have lasted at most several hundred million years. Thus, features on Alba Patera must have formed under atmospheric conditions essentially similar to the present and this epoch of channel formation was due to mechanisms other than the presence of a thick, warm atmosphere. Hydrothermal processes are a likely candidate.

In summary, the geological evidence for stable liquid water and the atmospheric models developed to explain this stability together suggest that conditions on Earth and Mars may have been fundamentally similar, and that this period of similarity may have lasted for several hundred million years and perhaps for as long as ~10^9 yr or so. These time scales are comparable with the amount of time it took for life to have originated on Earth. Clearly, planetary evolution led to very different histories for the two planets after this initial period of similarity. Our current understanding of planetary evolution would suggest that the root cause of the unfavorable (to life) turn of events on Mars was the incorporation of its atmospheric CO_2 into carbonate sediments. The accumulation of carbonates was a direct and inevitable result of Mars' small size and, hence, its inability to support and retain sufficient heat flow to power plate tectonic activity and thereby recycle the atmospheric constituents in a long-term geochemical cycle.

As atmospheric CO_2 was deposited as carbonates, lowering the pressure and temperature, any liquid water present would have become ice-covered. On Earth, ice-covered lakes are found in the Antarctic dry valleys and maintain liquid water under perennial ice covers despite a mean temperature of −20°C (see fig. 4-4). One biologically important feature of an ice-covered lake is its capacity to provide a thermally buffered habitat in the underice water column despite the cold external temperature. Consequently, microorganisms are capable of growing in these thermally stable lakes in regions that are otherwise bereft of life. By analogy, it is possible that ice-covered lakes on early Mars provided a relatively warm, liquid water environment for early Martian biota long after the surface temperature fell below freezing. Another feature of perennially ice-covered lakes is their ability to concentrate atmospheric gases in the water column. For example, Lake Hoare, Antarctica, has about three times the oxygen and 50% more nitrogen than would be in equilibrium with the atmosphere. Both biological and abiological processes contribute to the enhanced gas concentrations. Sedimentation and loss through the

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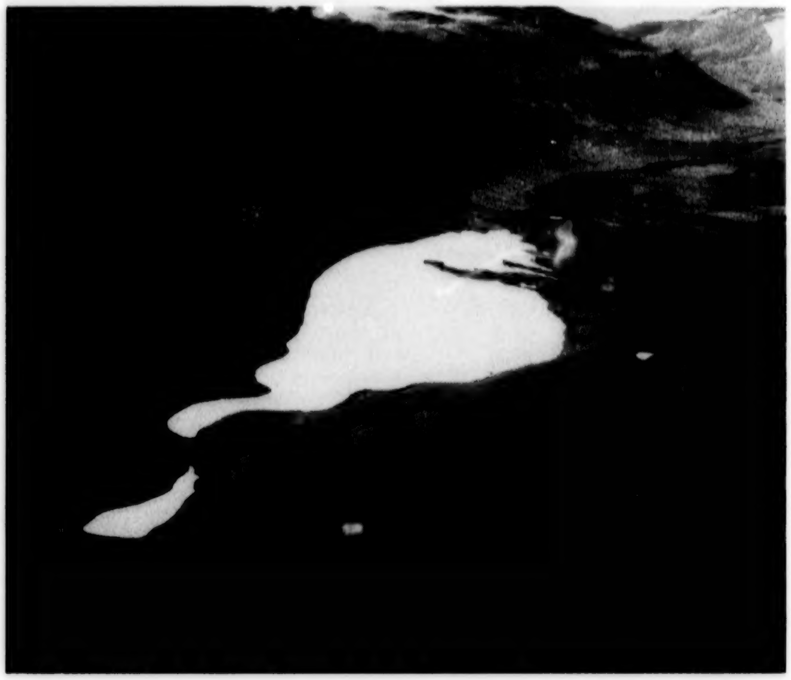

Figure 4-4. Lake Vanda, one of many perennially ice-covered lakes in the dry valleys of Antarctica. Despite mean temperatures of –20°C there is a persistent liquid water habitat under the ice cover of these lakes. The existence of ice-covered lake environments on early Mars may have considerably extended the period of potential biological activity after the mean surface temperature fell below freezing.

ice cover of organic carbon produced through photosynthesis represents a biological source of oxygen. Also, the incoming melt streams carry dissolved air into the lake. The gas is concentrated when the water freezes to the bottom of the ice cover. Both of these processes help to control the gas concentration in the lake water. These concentration mechanisms may have operated in the ice-covered Martian paleolakes, possibly enhancing the concentrations of biologically important gases (e.g., CO_2, N_2) from the thin Martian atmosphere. Thus as climatic conditions deteriorated on early Mars, ice-covered lakes could have been one of the last refuges—providing both thermal stability against a cooling external environment and enhanced gas concentrations of CO_2 and N_2 against a thinning atmosphere.

The formation of ice-covered lakes and the biota that may have inhabited these lakes is relevant to the question of continued carbonate precipitation on Mars after the mean surface temperatures fell below freezing. In lakes and shallow water environments on Earth, the presence of microorganisms accelerates the precipitation of carbonate by the removal of CO_2 from the local environment which lowers the pH. For this reason microbial mats and other benthic microflora are often encrusted in carbonate deposits. The presence or absence of organisms in Martian lakes could have been an important factor in setting the rate of carbonate deposition on early Mars.

Sediments deposited in lake environments may be identified from orbit or from the surface by the detection of stratified geological formations. The possible paleolake deposits on Mars detected from Viking orbiter imaging are therefore prime sites for a search for fossil evidence of past Martian life. The accumulation of detrital material and sediment on a lake bottom provides a fossil record spread over a large contiguous area. Although the shores may provide better sampling sites, there would be a good chance of detecting fossil remnants in sediments taken from anywhere on the lake bed. The possibility that carbonate formation was associated with aquatic environments on early Mars suggests that sites at which carbonate-bearing sediments are detected are prime locations for searching for organic material and/or biological fossils. Samples collected from depth could be analyzed for organic material and precipitates such as carbonates using pyrolysis techniques with analysis of the evolved gases.

Based upon the availability of liquid water on the Martian surface, the geological history of Mars can be divided into four epochs. In Epoch I, during which a primordial CO_2 atmosphere was actively maintained by impact and volcanic recycling, the mean annual temperature would have been above freezing, the pressure would have exceeded one atmosphere, and liquid water would have been widespread. Under such conditions, similar to early Earth, life could have arisen and become abundant. After this initial period of recycling, atmospheric CO_2 was irreversibly lost due to carbonate formation and the pressure and temperature declined. In Epoch II, the mean annual temperature fell below freezing, but peak temperatures would have exceeded freezing. As discussed above, ice-covered lakes, similar to those in the McMurdo Dry Valleys of Antarctica could have provided a habitat for life. In Epoch III, the mean and peak temperatures were below freezing and there would have been only transient liquid water. Microbial ecosystems living in endolithic rock "greenhouses" could have continued to survive. Finally, in Epoch IV, the pressure dropped to near the triple point pressure of water and liquid water could no longer have existed on the surface and life on the surface would have become extinct.

To explore the paleobiology of Mars many of the techniques and approaches that have been developed in the study of Earth's earliest biosphere can be applied. Finding evidence of past life on Mars will involve searching for many things, including direct traces of life such as microfossils, organically preserved cellular material, altered organic material, morphological microstructures and chemical discontinuities associated with life, isotopic signatures due to biochemical reactions, inorganic mineral deposits attributable to biomineralization, and hydrated minerals such as clays. Based upon our knowledge of fossils from Earth's Precambrian Era we can identify the types of targets that would be of paleobiological interest on future Mars missions. Figure 4-5 shows a diagram of the physical features that may be indicative of life in or near a lake environment. The sizes of these features range from micrometers to thousands of kilometers.

One of the most important forms of fossil evidence for microbial life on early Earth is stromatolites (fig. 4-6). These are defined as organosedimentary structures produced by sediment trapping, binding, and/or precipitation as a result of the growth and metabolism of microorganisms. Their formation is typically associated with the phototaxic properties of cyanobacteria and algae. The requirement for light drives these phototrophic organisms to move above any sediment that is deposited upon them leaving behind a layer of organic-rich sediment. In the absence of metazoan grazers, this sediment-building mechanism results in the formation of stromatolites. It is reasonable to suppose that if life arose on Mars, then the shallow water habitats would have been populated by algae-like photoautotrophs.

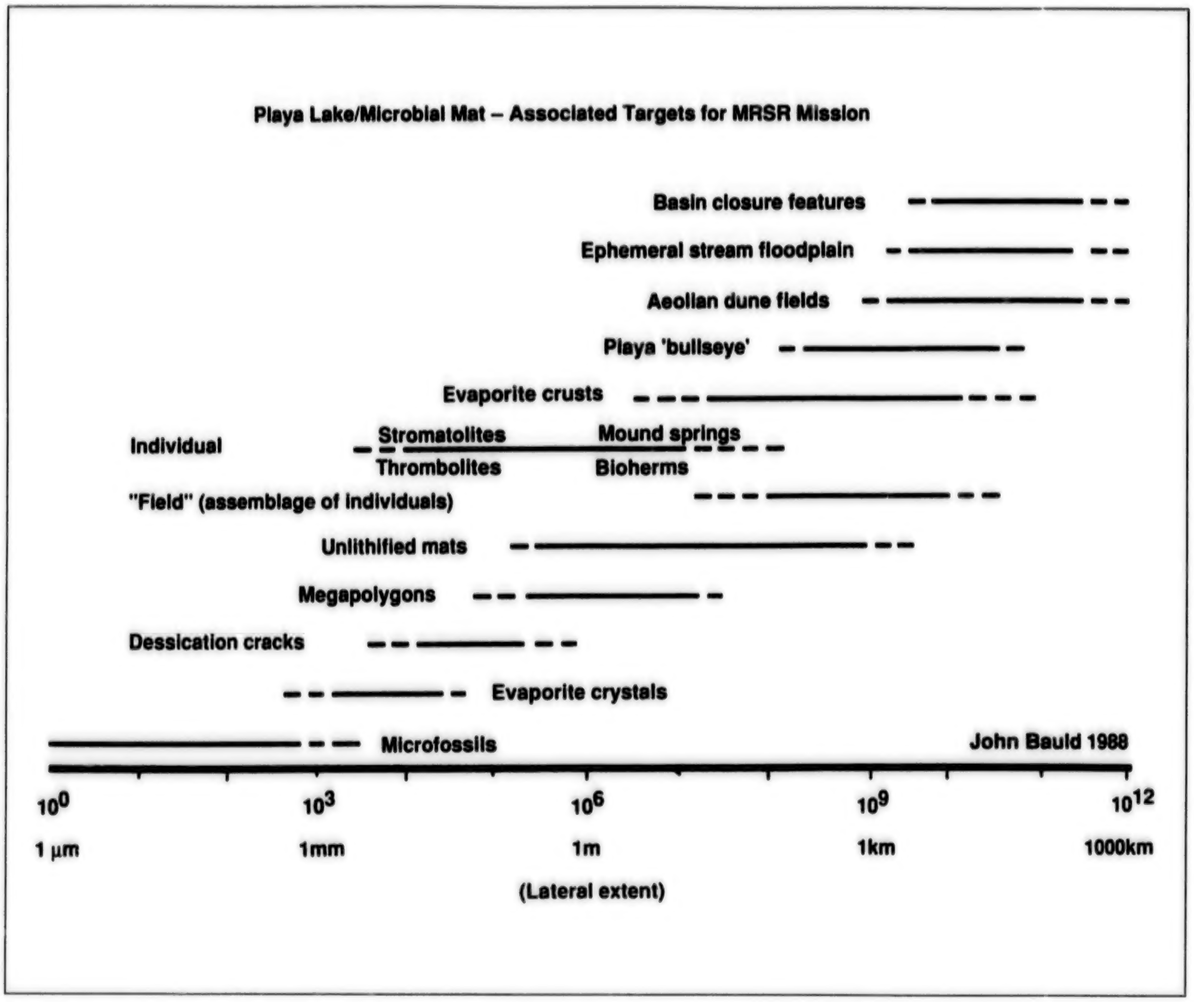

Figure 4-5. Approximate dimensions of biogenic and environmental targets diagnostic of, or consistent with, the former presence of saline/playa lake-hosted microbial mat communities on Mars. The search for fossil evidence of past life on Mars would begin with the investigation of such targets based upon orbital data.

Certainly it would appear that CO_2 was available on early Mars as a source of C atoms to any photoautotrophic organisms that may have existed to consume it. The prediction that stromatolites may be found on Mars is not just a reflection of our inevitably geocentric perspective on life. Rather, it springs from the fact that light is the most abundant source of energy on a planet and that motion toward light, through obscuring sediments, would have a selective advantage for those organisms that utilize it. Since stromatolites are macroscopic structures, often tens of meters in size, they are good targets for an *in situ* (and possibly remotely from orbit) search for fossil evidence of Mars' earliest biosphere.

Many of the techniques used to study early life on Earth may be applicable to the corresponding study on Mars. An interesting example of a powerful isotopic technique

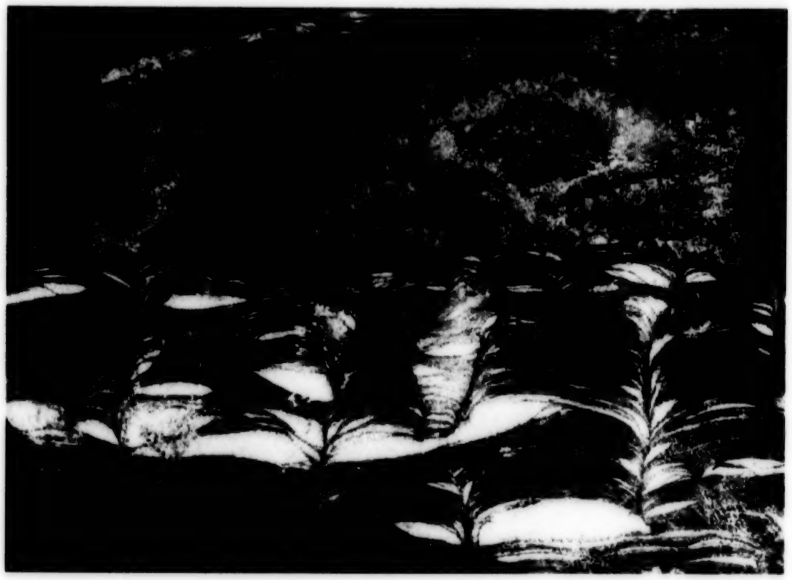

Figure 4-6. A picture of a precambrian stromatolite, 2.6 billion years old. Stromatolites are sedimentary structures formed as microorganisms move up through infalling sediments to reach the light. These macroscopic fossils of early life may be found on Mars. (Photo courtesy of M. Schidlowski.)

used on Earth is shown in figure 4-7. This plot shows the isotopic composition of carbon in sedimentary carbonate and organic material over most of Earth's history. Also shown are the isotopic shifts due to contemporary autotrophs. The approximately 2% shift in carbon isotopic composition in organic material is due to the selectivity of the carboxylase enzyme for the lighter isotope of carbon, ^{12}C. In studies of Earth's earliest biosphere this isotopic shift provides a useful test for determining if organic matter is of biological origin. It may be the case that similar effects may be present on Mars and might be a useful diagnostic of biological origin. Alternatively it may be possible that life on Mars exhibits no characteristic carbon isotope shift. Abiotic synthesis produces a variation of isotopic composition with carbon-number of the product and also varies with the method of synthesis. A uniform isotopic shift is more likely to be of biological origin.

Fossil evidence of early life on Mars may also be found frozen in the permafrost regions that are thought to cover both the northern and southern hemispheres of Mars poleward of about 40°. The most likely current location of the bulk of the waters on Mars is as permafrost in the polar latitudes. In addition to the above theoretical arguments for permafrost ice on Mars, there is some geomorphological evidence as well. The equatorial regions on Mars appear to be depleted in ground ice while evidence for ground ice, in the form of quasi-viscous relaxation of topography due to creep deformation of ice, is widespread at high latitudes. Ground ice can exist in equilibrium with the atmosphere only at those latitudes and depths where crustal temperatures are below the frost point. Outside these regions, ground ice can survive only if it is diffusively isolated from the atmosphere by a regolith of low gaseous permeability.

The age of the Martian permafrost is uncertain, but it can be broadly classified, based upon crater densities. In the northern hemisphere there are plains with low crater densities which seems to indicate that they are of recent age (younger than 2 billion years old). In the southern hemisphere the terrain is heavily cratered, essentially dating back to the end of the late heavy bombardment some 3.8 billion years ago (the stratigraphic period designated as the Noachian).

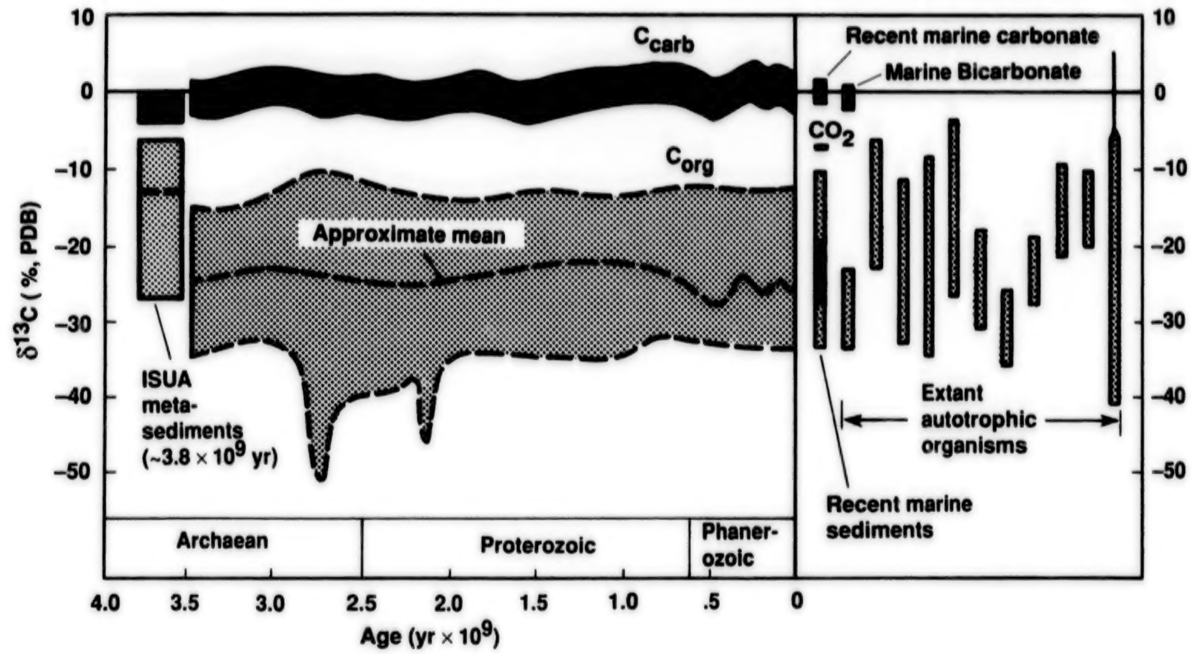

Figure 4-7. Carbon isotopic values for carbonate and organic sediments on Earth for the last 3.8 billion years. Carbon isotope values are reported in units of ‰ which corresponds to the enrichment of the heavy isotope ^{13}C in parts per thousand, when compared to a standard. Photosynthetic systems preferentially uptake ^{12}C over ^{13}C resulting in a isotopic shift of about −25 ‰. This isotopic shift has characterized organic sediments throughout Earth's history despite significant changes in the Earth's environment and biota. Similar carbon isotope values in ancient Martian organic material may be an indicator of biological origins.

As is the case on Earth, even as mean global temperatures on Mars were above freezing, polar permafrost could form and persist over geological times. Thus, material deposited, and preserved in the ancient permafrost regions of the southern hemisphere, should record events (and biota if any) that existed on Mars during the time that global conditions may have been suitable for life.

Material trapped within the permafrost of Mars should be exceptionally well preserved. The mean annual temperature of the permafrost regions is less than 200 K (−73°C) and material below several meters of depth would be well shielded from the cosmic radiation environment. The Martian climate of the past 3 billion years, while unsuited for life, is ideal for the preservation of frozen samples. In addition, the low erosion and burial rates (estimated to be about 1 meter/billion years) and the virtual absence of plate tectonics imply that the ancient material is accessible from the surface and has not been altered by internal heat or tectonic activity. Thus, we suggest that Martian permafrost may provide an ideal target for future exobiological investigations of Mars.

The discovery of fossil evidence of past life on Mars would have significant implications—so, too, would the determination that life did not originate on Mars. Our understanding of the processes that lead to the origin of life on Earth are uncertain at best and we do not know if life is a singular event or if it is widespread in the universe. Some theories of the formation of life in the cosmos would suggest that life is widespread. Finding evidence for past life on Mars would be a crucial confirmation of that hypothesis—the first data point other than Earth life. If detailed investigations of sediments on Mars confirm the view that there was a similarity between early Earth and early Mars, and yet life did not evolve on Mars, it may be an indication that the origin of life is a far more exacting process that we currently envision and is therefore far less widespread in the cosmos.

If life did evolve on Mars, as it did on the Earth, a better record of these events may be preserved in the sediments of Mars than is available on Earth. On Earth unaltered rocks that date back to the 3.5–4.0-billion-year-old timeframe are rare, while on Mars over half of the planet (primarily the southern hemisphere) dates back to the end of the late bombardment, some 3.8 billion years ago. Thus, while there may be no life on Mars today, it may hold the best record of the events that led to the origin of life on Earth-like planets. Mars still holds much interest from a biological perspective.

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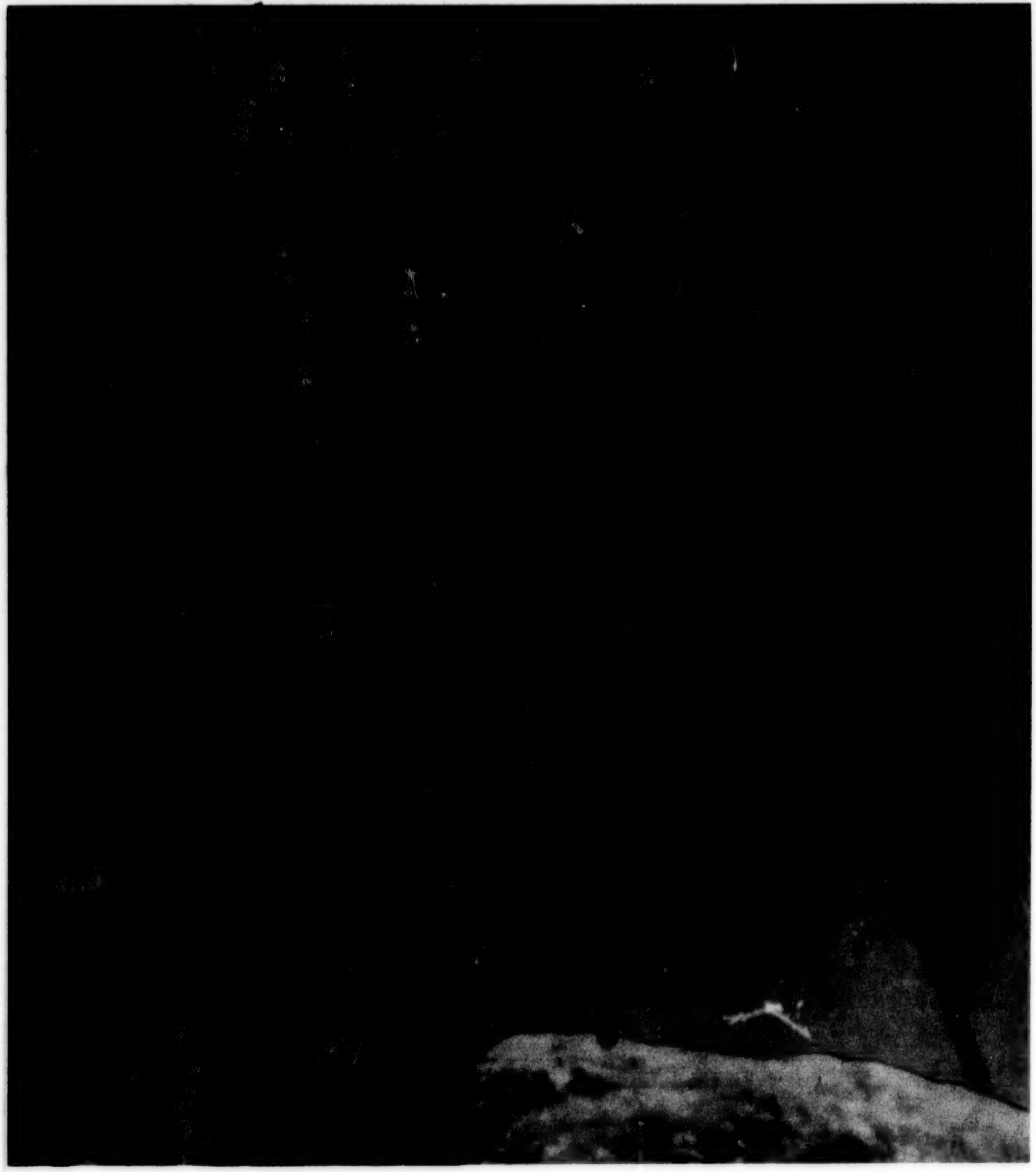

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Giant Planets: Clues on Current and Past Organic Chemistry in the Outer Solar System

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The giant planets of the outer solar system—Jupiter, Saturn, Uranus, and Neptune—were formed in the same flattened disk of gas and dust, the solar nebula, as the terrestrial planets—Mercury, Venus, Earth, and Mars—were. Yet, the giant planets differ in some very fundamental ways from the terrestrial planets. As the name would indicate, the giant planets are both bigger and more massive than the terrestrial planets. The sizes of the giant planets range from about 4 to about 11 times that of the Earth and their masses from about 15 to 300 times that of the Earth. Hydrogen and helium are the most abundant elements in the atmospheres of the giant planets, just as in the Sun. However, nitrogen, oxygen, and carbon are the most abundant elements in the atmospheres of Venus, Earth, and Mars, the three terrestrial planets with significant atmospheres.

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Furthermore, the atmospheres of the giant planets extend about ten thousand to several tens of thousands kilometers into the interiors of the giant planets, thereby representing a significant fraction of the entire mass of these planets, whereas the atmospheres of the terrestrial planets have much smaller depths and constitute only a tiny fraction of these planets' masses. Finally, numerous moons and rings surround each of the giant planets, whereas the terrestrial planets have no rings and at most two moons (Mars).

Despite these enormous differences, the giant planets are relevant to exobiology in general and the origin of life on the Earth in particular. The chemical steps that led to the first genetically reproducing organisms are widely believed to have involved the production of complex carbon-containing molecules from simpler carbon-containing molecules that occurred in a reducing environment (little free oxygen was present). The atmospheres of the giant planets represent highly reducing environments, in which one of the simplest carbon-containing molecules—methane—is being converted into more complicated molecules. Thus, these atmospheres represent natural laboratories for observing and understanding some of the initial chemistry that leads to the production of organic molecules. Furthermore, the compositions of the atmospheres of the giant planets provide important constraints on the composition and abundance of carbon-containing molecules in the outer part of the solar nebula which, therefore, lead to insights on possible sources of carbon and other biologically relevant materials for the origin of life on Earth. Finally, if these materials came to Earth from the outer solar system, as they may have, the giant planets played a fundamental role in transferring this material to the inner solar system by virtue of their gravitational interactions with small bodies.

This chapter begins with the giant planets as they are today, and discusses their basic properties and the chemistry that is occurring in their atmospheres. Then, the chapter explores theories of their origin and stress aspects of these theories that may have relevance to exobiology and the origin of life on Earth.

Properties of the Giant Planets

The giant planets are made of three basic materials: gas, ice, and rock. Gas refers to compounds that are made almost exclusively of the elements hydrogen and helium; ice refers to those containing mixtures of water and carbon- and nitrogen-containing materials; and rock refers to ones made of mixtures of silicon, magnesium, iron, oxygen, and other heavy elements. This separation reflects the differing abilities of materials in these three classes to condense in different parts of the solar nebula. The gas component did not freeze out anywhere in this nebula; the ices, especially water, condensed in the outer part of the nebula, where the giant planets formed; and rock also condensed in the inner part of the solar nebula, where the terrestrial planets formed. Thus, the terrestrial planets are composed almost exclusively of rock, whereas the giant planets are made of varying proportions of all three materials. These names are merely meant to define compositional classes. The deep interiors of the giant planets are hot enough to melt and vaporize rock and ice-containing compounds that are located in them.

Figure 5-1 illustrates the interior structures of the giant planets that have been derived by fitting physically based models to such properties as their masses, sizes, and mass distributions (or more precisely, their gravitational moments). The outer shells or envelopes of the giant planets are composed primarily of the gas component. Elemental hydrogen occurs as molecular hydrogen in the outer part of these envelopes. However, sufficient pressures are reached in the deeper parts of the envelopes of Jupiter and Saturn (pressures in excess of several million times that at the Earth's surface) for molecular hydrogen to be pressure dissociated and ionized into a highly electrically conducting form, metallic hydrogen. It is within these regions that the strong magnetic fields of Jupiter and Saturn are produced.

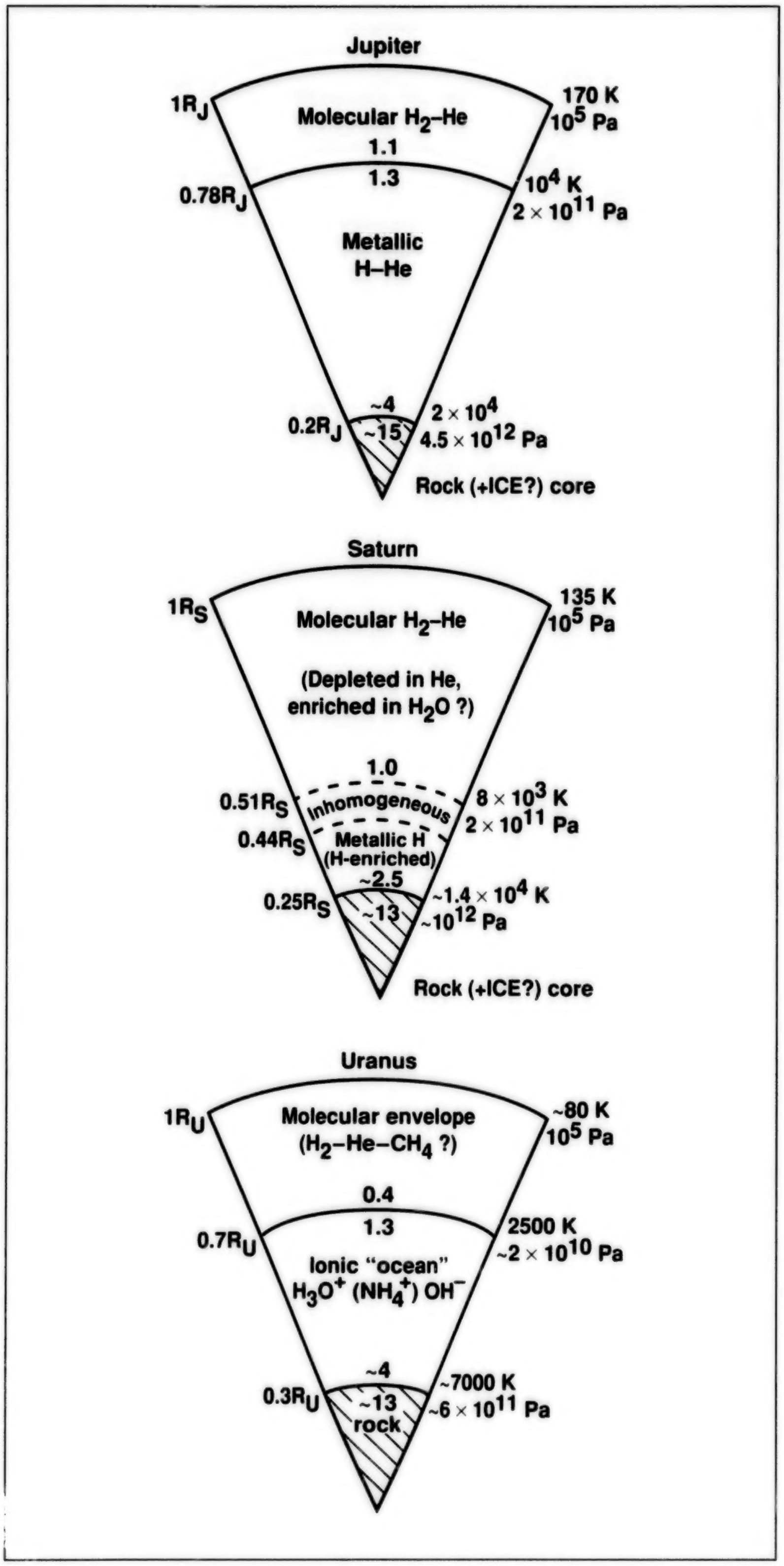

Jupiter
1R_J
Molecular H2–He
170 K
10^5 Pa
1.1
1.3
0.78R_J
10^4 K
2 × 10^11 Pa
Metallic H–He
~4
~15
0.2R_J
2 × 10^4
4.5 × 10^12 Pa
Rock (+ICE?) core
Saturn
1R_S
Molecular H2–He
135 K
10^5 Pa
(Depleted in He, enriched in H2O ?)
1.0
0.51R_S
Inhomogeneous
8 × 10^3 K
2 × 10^11 Pa
0.44R_S
Metallic H (H-enriched)
~2.5
0.25R_S
~13
~1.4 × 10^4 K
~10^12 Pa
Rock (+ICE?) core
Uranus
1R_U
Molecular envelope (H2–He–CH4 ?)
~80 K
10^5 Pa
0.4
1.3
0.7R_U
2500 K
~2 × 10^10 Pa
Ionic "ocean" H3O+ (NH4+) OH−
~4
0.3R_U
~13
rock
~7000 K
~6 × 10^11 Pa

Figure 5-1. Schematic representation of the interior structure of the giant planets. Neptune has a structure similar to that of Uranus and so it is not shown. The numbers on the left show distance from the center in units of the planet's radius; the numbers on the right show pressure and temperature in units of pascals (=10^{-5} bars) and degrees Kelvin, respectively; and the numbers in the center show density in units of that of liquid water.

Table 5-1: Properties of the Giant Planets

Property	Planet			
	Jupiter	Saturn	Uranus	Neptune
Distance from Sun (Earth's distance = 1)	5.20	9.55	19.22	30.11
Radius (Earth's radius = 1)	11.27	9.44	4.10	3.88
Mean density (water = 1)	1.31	0.69	1.19	1.66
Total mass (Earth's mass = 1)	318.1	95.1	14.6	17.2
Mass of gas component	254–292	72–79	1.3–3.6	0.7–3.2
Mass of rock and ice	26–64	16–23	11–13.3	14–16.5
Axial inclination (degrees from normal to orbital plane)	3.1	26.7	98.0	29
C/H (solar = 1)	2.3	5.1	35	40

Near the central regions or cores of the giant planets, ice and rock are the dominant materials. It used to be thought that ice and rock were totally or almost totally segregated from the gas parts of the giant planets. However, there is an increasing amount of evidence indicating that some mixing of these materials has occurred. Thus, the envelopes contain a significant fraction of the planets' bulk content of rock and ices and the cores, at least in the case of Uranus and Neptune, contain some gas.

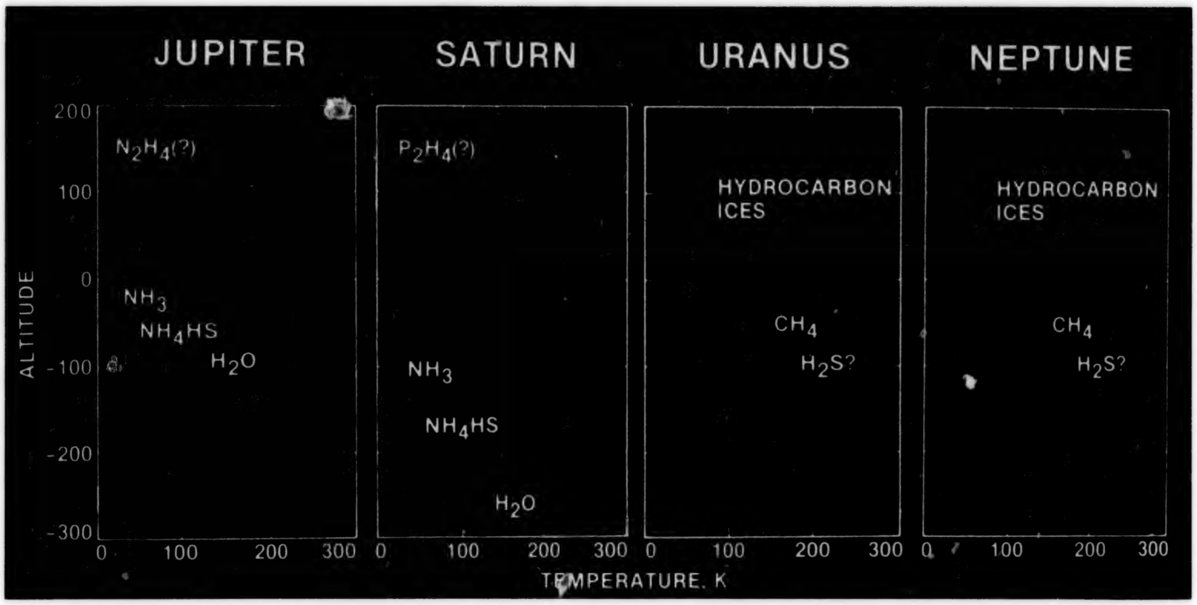
The giant planets are not simply different sized versions of the same thing, but exhibit a wide variation in their basic properties, as summarized in table 5-1. Not only do their masses and sizes vary considerably, but so does the relative proportion of gas and ice plus rock. It is currently not possible to unambiguously separate the rock and ice components. Massive Jupiter is made mostly of gas, whereas less massive Uranus and Neptune are composed primarily of ice and rock. Saturn is made primarily of gas, but has a larger fractional abundance of ice and rock than does Jupiter. An important point for later consideration is that the masses of the rock plus ice components of the four giant planets are the same within a factor of several, whereas the masses of the gas component vary by a factor of 100.

The very outer portion of the envelopes of the giant planets, their atmospheres, is the part that can be photographed and directly sensed with a wide variety of instruments. As shown in the chapter frontispiece, there is a wide diversity in the appearances of these observable atmospheres, with Jupiter showing an incredible range of structures and colors and Uranus being almost featureless. This diversity reflects a combination of the densities and locations of the cloud layers in these atmospheres, the strength of the atmospheric motions, and the nature of the coloring agents that are produced in these atmospheres.

It is convenient to divide these atmospheres into two major regions, by analogy to a similar division for the Earth's atmosphere. Within the deeper lying troposphere, temperatures increase continually with increasing depth, whereas the reverse is true of the stratosphere. The temperature inversion of the Earth's stratosphere is due, in part, to the absorption of sunlight by ozone, whereas the inversion of the outer planets' stratospheres is caused by sunlight being absorbed by gaseous methane and small particles or aerosols. In both cases, the stratosphere's ability to cool by emitting thermal radiation diminishes with increasing altitude (decreasing density) and this abets the production of the inversion. Below the uppermost parts of the troposphere, temperatures in all four atmospheres increase as rapidly as is physically possible, at the adiabatic lapse rate. This is the rate established when there is efficient convection. Heat released from the deep interiors of the giant planets helps to drive the convection. Indeed, about half the energy radiated by Jupiter, Saturn, and Neptune is derived from internal heat, with the rest coming from absorbed sunlight. Internal heat contributes a small fraction (as yet unmeasured at the time this paper was presented) to the thermal radiation emitted by Uranus. By contrast, all but a tiny fraction of the radiation the Earth emits to space is due to absorbed sunlight.

Extremely low temperatures occur at the tropopause, the boundary between the stratosphere and the troposphere. Values of approximately 50, 50, 80, and 100 K characterize the tropopauses of Neptune, Uranus, Saturn, and Jupiter, respectively. As a result of the progressively lower temperatures with increasing altitude in the upper tropospheres of the giant planets, a series of ice condensation cloud layers form. In the cases of Jupiter and Saturn, first water clouds, then ammonium hydrosulfide clouds (formed from hydrogen sulfide and ammonia gases), and finally ammonia clouds form as the tropopause is approached (fig. 5-2). In the cases of Uranus and Neptune, even methane condenses in the upper troposphere, where it forms the highest cloud layer. The next cloud down in these atmospheres may be made of hydrogen sulfide, rather than ammonia ice, due to differing proportions of hydrogen sulfide and ammonia in these planets' atmospheres as compared to those of Jupiter and Saturn.

Figure 5-2. Schematic structure of the atmospheres of the giant planets.



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Chemistry

If an outer planet atmosphere was in local thermodynamic equilibrium, only the most stable molecules and atoms would be present. In this case, methane (CH_4) would be essentially the only carbon-bearing molecule and ammonia (NH_3) would be the only nitrogen-bearing molecule at the relatively low temperatures and moderate pressures that characterize these atmospheres. However, small amounts of molecules having more than one carbon atom, such as ethane (C_2H_6) and acetylene (C_2H_2), have been detected in the stratospheres of these planetary atmospheres, and small quantities of carbon monoxide (CO) and hydrogen cyanide (HCN) have been detected in Jupiter's troposphere. Even such exotic molecules as germane (GeH_4—four hydrogen atoms combined with a single atom of germanium) and phosphine (PH_3—three hydrogen atoms combined with a single atom of phosphorus) have been detected in the tropospheres of Jupiter and Saturn.

Marked departures from local thermodynamic equilibrium, as illustrated by the presence of the above species, arise from a combination of local energy sources producing them from the more stable compounds and from vertical motions carrying molecules from the deep interior, where a different thermodynamic state occurs, to the observable atmosphere. Key energy sources for driving atmospheric chemistry include solar UV radiation, lightning, and high energy charged particles (protons, heavy ions, and electrons) that are precipitated into the auroral zones from the Van Allen belts, or magnetospheres, surrounding these planets.

Despite the fact that the amount of solar energy reaching the giant planets is small—at Jupiter, only 4%, and at Neptune, a meager 0.1%, of that at the Earth—the chemistry of methane, and, for that matter, of virtually all constituents in their atmospheres, is controlled largely by the available solar flux. This is evident from the relative strengths of the various energy sources at Jupiter, listed in table 5-2. The UV portion of the solar energy at Jupiter amounts to just 1% of the total; it is, nevertheless, this radiation which is responsible for initiating the photochemical processes. The photons or quanta of light at UV wavelengths are sufficiently energetic to break the chemical bonds of stable molecules and therefore are the ones of interest for atmospheric chemistry.

Lightning can be potentially important in producing certain disequilibrium species such as HCN and CO. Energetically charged particles

Table 5-2: Energy Sources for Chemistry at Jupiter

Solar energy	
Total incident, F_T	5×10^4 erg cm^{-2} s^{-1}
Ultraviolet, F_{UV}	~1% of F_T
Lightning (electric currents) + Thunder (acoustic waves)	0.001% of F_T (terrestrial), could be as high as 0.003% of F_T at Jupiter
Magnetospheric (auroral)	
Total input	10^{13} W, or 10 erg cm^{-2} s^{-1} 0.02% of F_T
Globally averaged	0.4 erg cm^{-2} s^{-1}, or 0.001% of F_T

C-2

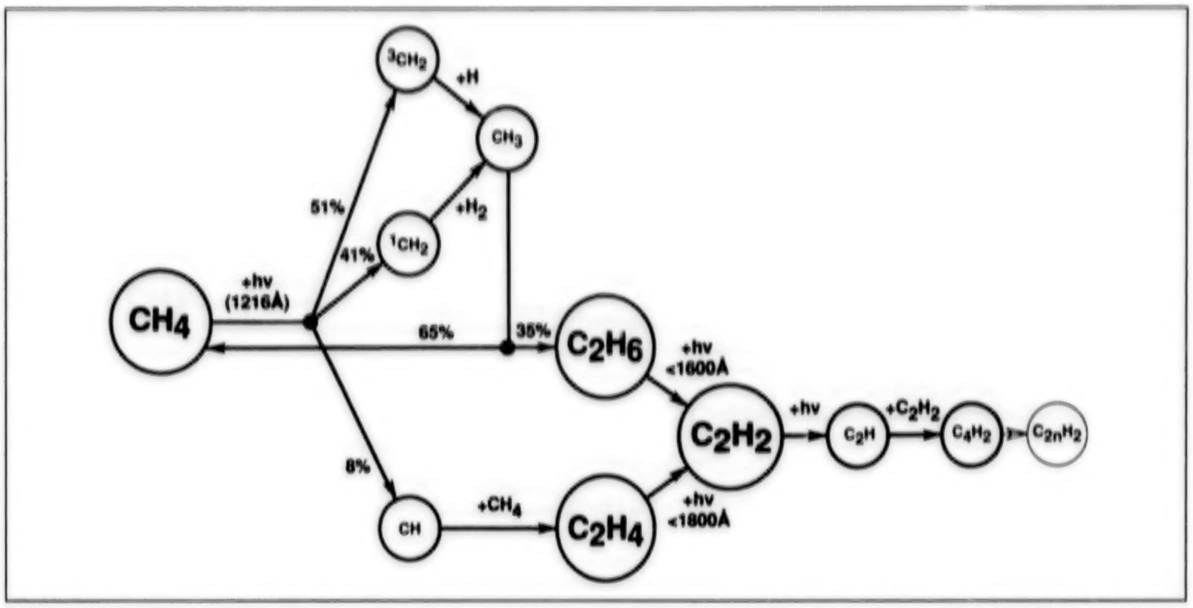

Figure 5-3. Schematic of the principal pathways for methane photochemistry in the atmospheres of the giant planets.

may play a role in the chemistry at high latitudes, where the auroral zones occur, perhaps resulting in the production of some heavier hydrocarbons and small particles (aerosols). The photochemistry of methane is discussed first in the following paragraphs, followed by a discussion of the effects of other energy sources.

The most significant pathways in the photochemical processes of CH_4 at Jupiter are shown schematically in figure 5-3. Although the photochemistry of methane is initiated by absorption of solar photons with wavelengths below 1450 Å, for all practical purposes it is the extremely large solar flux at the Lyman-alpha wavelength of 1216 Å (a strong line of the abundant hydrogen atoms in the Sun) which is responsible for more than 90% of the bond-breaking or dissociation of CH_4 into molecular and atomic fragments. These fragments or radicals are very reactive and combine with themselves and other molecules to produce more stable compounds.

The photodissociation of CH_4 by the solar Lyman α line of hydrogen produces the CH_2 radical in its singlet (1) and triplet (3) states in roughly equal proportions. These states refer to the number of closely spaced energy levels characterizing the radical's lowest energy level. A small fraction (8%) of the CH radical is also produced. Note the conspicuous absence of CH_3, as it is kinetically forbidden as a CH_4 photodissociation product. Both states of CH_2 in turn produce the methyl radical, CH_3, on reaction with molecular hydrogen. The reaction of

CH_3 with itself produces C_2H_6; whereas the reaction of CH_3 with atomic hydrogen, H, recycles or regenerates methane. Nearly two-thirds of methane is recycled in this manner, thereby leading to no net production of more complicated molecules. The reaction of CH with CH_4 produces ethylene, C_2H_4. C_2H_4 and C_2H_6 photodissociate below 1800 Å and 1600 Å, respectively, to produce C_2H_2.

Ethane is highly stable in the atmosphere as it is largely shielded from the dissociative UV flux by the much more abundant methane, which also absorbs photons in the same wavelength range as ethane. Moreover, the photolysis rate of C_2H_6 is only 10% that of C_2H_4. Likewise C_2H_2 is also stable due to the low quantum efficiency (molecules produced per quantum of light absorbed) for its photolysis products, as well as rapid recycling of these products back to C_2H_2. The photodissociation of C_2H_2 proceeds slowly, resulting eventually in the formation of diacetylene (C_4H_2) and higher order polyacetylenes. In addition to the above mentioned hydrocarbon species, methane photochemistry is expected to produce other higher order hydrocarbons, such as propane (C_3H_8), methylacetylene (C_3H_4), butane (C_4H_{10}), and even benzene (C_6H_6), a molecule with a ring structure. Thus, UV sunlight acting on methane in the stratospheres of the giant planets produces small quantities of ethane and acetylene, along with smaller amounts of more complex carbon compounds.

The fate of methane in the photochemical processes on the major planets is its conversion to heavier hydrocarbons with multiple carbon atoms. Eventually these heavier hydrocarbons would be removed from the stratosphere by condensation followed by rain-out or snow-out. In the deeper and warmer parts of the troposphere, the evaporation and thermal decomposition of these hydrocarbons, followed by high temperature-high pressure chemistry in the presence of molecular hydrogen, would regenerate methane. The latter would be convected up to the higher atmosphere, thus stabilizing this constituent on the major planets. In the absence of these and the stratospheric recycling mechanisms, all of the methane on Jupiter would have been irreversibly destroyed in a few thousand years. Its present day abundance on the major planets attests to the effectiveness of these recycling processes.

Disequilibrium species, such as HCN, CO, PH_3, and GeH_4, have been detected in the visible atmosphere of Jupiter. Some have been seen on Saturn as well. None can be produced by conventional photochemical processes. The following paragraphs discuss some hypotheses for explaining their presence, and their significance.

HCN is an important precursor molecule for the formation of amino acids and proteins, which are important molecules in living organisms. The detection of HCN in the atmosphere of Jupiter poses a dilemma, as photochemical reactions fail to produce its observed abundance (2 parts per billion). Nevertheless, it is instructive to review these processes since the lack of complete information on some key reaction rates still makes them potentially important.

The photolysis of methylamine (CH_3NH_2) or ethyleneimine (C_2H_5N) is expected to produce HCN. These precursor molecules are produced by the coupling of the NH_3^- and CH_4^- photochemistries in a relatively narrow altitude range of the upper troposphere. The quantum yield of HCN from the photolysis of CH_3NH_2 is low. It is therefore not an important source of the observed quantity of HCN in the Jovian atmosphere. In the second case (C_2H_5N source), the rate of the reaction between NH_2 and C_2H_3 (which are intermediate products of the NH_3 and CH_4 photochemistries) that helps to produce it is not known; neither are the reaction kinetics of all the possible products. Thus, the photochemical source of HCN is, at best, speculative and only potentially important.

Another mechanism proposed for the formation of HCN is lightning. This source is also highly controversial. Lightning discharges have been detected on long exposure images of the Jovian atmosphere obtained by the Voyager spacecraft. Lightning discharges produce HCN in a methane/ammonia atmosphere by temporarily raising the local atmosphere to very high temperatures, where reactions can proceed rapidly, and by producing UV radiation. The possible importance of lightning for generating HCN and other disequilibrium species in Jupiter's atmosphere depends strongly on what assumptions are made about the fraction of the available energy in the atmosphere (that due to sunlight and heat convected from the interior) that is converted into lightning and the location of the lightning. For example, its occurrence near or within tropospheric water clouds could abet the production of some disequilibrium species, including CO as well as HCN.

Disequilibrium species such as PH_3, GeH_4, and CO are the major thermodynamical equilibrium forms of P, Ge, and C, respectively, in the unobservable, very deep, high temperature and pressure regions of the giant planets' tropospheres. Yet they have been detected in the observable atmosphere of Jupiter (PH_3 and CO have been detected in Saturn's stratosphere also). Their presence in these atmospheres implies that either strong vertical mixing is transporting these compounds from great depths into the upper troposphere and/or stratosphere or that an extraplanetary source is introducing significant quantities of the needed atoms (e.g., O for CO) into the atmosphere. The latter mechanism might be an important source for CO. Influx of oxygen-bearing species, such as water from infalling meteorites, comets, ring particles, and material derived from the planets' moons, could initiate a series of reactions with atmospheric hydrocarbons, which would eventually produce CO. However, recent measurements indicate that the fractional abundance of CO is higher in the troposphere than the stratosphere and, therefore, favor an internal source.

If the disequilibrium gases, GeH_4, PH_3, and CO, are derived from the deep tropospheres of Jupiter and Saturn, they need to undergo strong vertical mixing to bring them to the observable atmospheres before they can be entirely converted to their low-temperature, thermodynamic equilibrium forms. Since the rates at which this conversion occurs decrease very rapidly with decreasing temperature (and hence increasing altitude), the abundance of these disequilibrium gases in the atmosphere is essentially set by their abundances at a "quench" temperature, where the transport and reaction rates are comparable. The quench temperature is the lowest value at which thermodynamic equilibrium is readily achieved.

The above mentioned mixing hypothesis is attractive from another viewpoint—that for explaining the color of the Great Red Spot (GRS) on Jupiter and other cloud features. (Other possibilities are noted in the next paragraph.) If PH_3 is indeed mixed up to the upper troposphere by strong upward transport, it is likely to undergo UV photolysis which could eventually yield triclinic red phosphorus crystals, P_4(s), in the Jovian atmosphere. This chromophore resembles the color in the GRS and many other clouds on Jupiter and Saturn. Since the GRS may be a region of enhanced vertical motion, a larger amount of PH_3 might persist to near the top of the GRS clouds before undergoing chemical conversion to P_4(s), thereby producing a particularly strong coloration of the GRS.

Other possible candidates suggested for explaining these cloud colors include H_2S and some polyacetylenes. Generally polyacetylenes are white; however, some as yet poorly understood chemistry could produce the appropriate color from these hydrocarbons. For example, solar UV radiation acting on solid polyacetylenes may produce more complicated forms that do absorb at visible wavelengths.

The photolysis of H_2S occurs below 3000 Å, and eventually produces elemental sulfur, S_8, which can be yellow, or hydrogen polysulfide, H_xS_y (x and y are various integers), which is orange, or ammonium polysulfide $[(NH_4)_xS_y]$, which is brown. Thus a wide ranging choice of colors is possible as a result of H_2S photochemistry. This is attractive, as the colors of the Jovian clouds also range from yellowish red to orange to brown to red. The principal difficulty with the scenario is that the fate of H_2S on Jupiter and Saturn is most likely its removal by ammonium hydrosulfide (NH_4SH) cloud condensation before it could get a chance to be photolyzed (NH_4SH is formed by the reaction of NH_3 with H_2S). H_2S has not been detected in the upper troposphere of Jupiter (or any other giant planet). The upper limit on its fractional abundance at the 700 millibar level is 0.001 times the amount expected from the solar abundances of S and H. These observations seem consistent with its loss in the NH_4SH cloud just discussed. Furthermore, because of a severe depletion due to scattering by Jovian air molecules, the 3000 Å and shorter wavelength photons that are needed for H_2S photolysis cannot penetrate to the level below these clouds (pressures greater than 2 bars) where it is available.

Another possibility is that, like CO, sulfur in some form might be brought into the upper atmosphere from an extraplanetary source. For Jupiter, such a source would presumably be the circum-Jovian Io plasma torus which is populated by sulfur ions derived from the explosive volcanoes of the moon Io. The incoming sulfur would react with the constituents in the Jovian upper atmosphere, producing perhaps CS, and even COS if oxygen is also being injected at the same time as sulfur. None of these major sulfur compounds have the color that matches the color in the clouds of Jupiter. A possibility exists that subsequent chemistry induced by energetic charged particles could produce trace species with the right color.

Sufficiently energetic charged particles, namely cosmic rays, penetrate to the deep troposphere where they could interact with H_2S or NH_4SH to produce species (such as S_8, H_xS_y, etc.) which have the colors of the clouds of Jupiter and Saturn. This would circumvent the difficulty associated with photolyzing H_2S. It would also apply equally to Jupiter and Saturn. Much laboratory work needs to be done before one can fully understand the implications of the charged-particle-induced chemistry on the giant planets. In any event, the striking and varying coloration of the clouds of Jupiter and Saturn strongly suggest the occurrence of chemical processes that produce complex molecules.

Origin of the Giant Planets

The composition of the giant planets provides useful clues and constraints on the manner in which they formed. First, they all contain large amounts of the gas component. This means that growth to their current masses was completed or almost completed before the gases of the solar nebula were dissipated. Based on the properties of young stars of comparable masses to that of the Sun, the T Tauri stars, the time scale for the formation of the giant planets was ten million years or less. Also, an efficient means of concentrating the nebula gases around the giant planets is required.

All the giant planets have a much larger fraction of the elements found in the ice and rock components, relative to hydrogen, than does the Sun. The degree of this heavy element excess varies from about a factor of 5 for Jupiter to 1000 for Uranus and Neptune. Thus, the giant planets accreted the solid phase of the solar nebula much more efficiently than its gas phase. This property suggests that the giant planets grew initially in the same way that the terrestrial planets did, but that they followed their own unique path of formation at a later stage.

Another important constraint on the formation of the giant planets is the similarity of their rock and ice masses and the dissimilarity of their gas masses. This constraint may imply that there is a rather abrupt transition in the accretion of the giant planets in which they switch from a solid-dominated accretion to a gas-dominated accretion. Furthermore, this transition point may have occurred at similar rock and ice masses.

Finally, the amount of carbon in the form of methane in the atmospheres of the giant planets, relative to hydrogen, is more than is found in the Sun. In particular, the carbon-to-hydrogen ratios in the atmospheres of Jupiter, Saturn, Uranus, and Neptune are about 2, 5, 35, and 40 times larger than the corresponding solar ratio and presumably that of the solar nebula (counting both solids and gases). This enhancement of carbon implies that some carbon was contained in the solid phases of the outer solar nebula (remembering that the giant planets preferentially accreted the solids) and that some of the carbon in the solid phase was mixed into the envelope.

The above considerations have led to the following hypothesis for the origins of the giant planets. The initial accretion of each of the giant planets was essentially identical to that of the terrestrial planets. Through a large number of gentle collisions, small solid bodies grew to successively larger sizes. In a given collision between two bodies, some of the kinetic energy of their relative motion before impact is dissipated upon impact, i.e., it is converted into other forms of energy, such as heat, sound waves, and fracturing. When the remaining kinetic energy is less than the energy due to either their mutual gravitational attraction or surface sticking forces, a composite body results. In a given region of the solar system, the largest of these growing bodies eventually interacted with and usually accreted all the smaller bodies or planetesimals, whose orbits crossed its orbit. Thus, ultimately the biggest accreting bodies attained masses comparable to or greater than that of the present Earth.

Once the masses of the forming giant planets became large enough, they were able to begin gravitationally concentrating large amounts of gas from the surrounding solar nebula about themselves. The gas within a forming planet's sphere of influence cooled by emitting thermal radiation to space and, as a result, the gaseous envelope contracted. This caused more gas to be added to the sphere of influence. Initially, the envelope's contraction rate was very slow and so the planet's mass was dominated by the solid planetesimals it had accreted. However, as the planet's mass approached ten to several tens of Earth masses, the rate of the envelope's contraction dramatically accelerated and soon a point was reached where accretion was dominated by the addition of gas from the surrounding solar nebula. This highly nonlinear relationship between the masses of the envelope and core is illustrated in figure 5-4. Once the envelope and core masses became comparable, subsequent growth was dominated by the gas component. This point of mass equality is sometimes referred to as the critical core mass, since a giant planet needed to attain this mass to add very large amounts of gas subsequently.

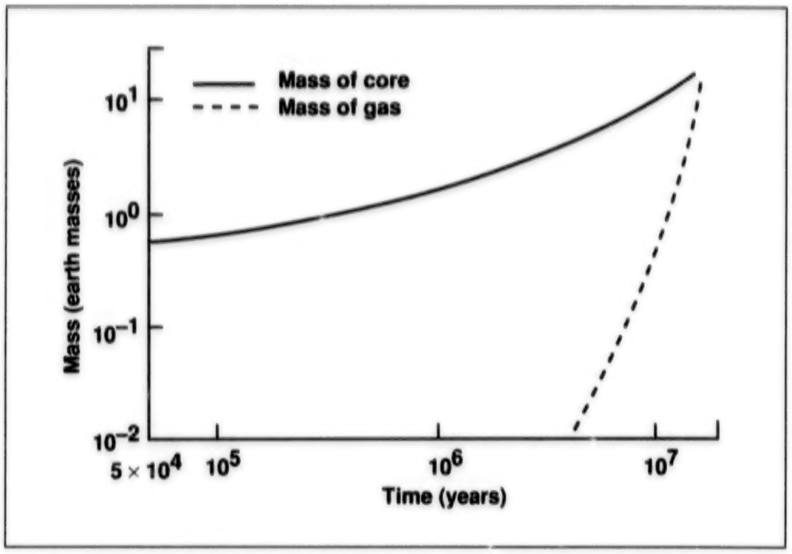

Figure 5-4. Envelope mass (dashed) and core mass (solid) as a function of time for a model of the growth of a giant planet for which the planetesimals were accreted at a constant rate of 10^{-6} Earth masses per year.

Simulations of the growth of the giant planets indicate that the value of the critical core mass is very insensitive to the pressure and temperature of the surrounding solar nebula, i.e., the distance from the Sun at which the planet formed. The critical core mass depends most sensitively on the rate of accretion of the solid planetesimals, with larger accretion rates implying a somewhat larger value. If we use the masses of the ice and rock component of the giant planets as a guide to the values of the critical core masses, we find that they should have formed in times comparable to or less than reasonable lifetimes for the solar nebula. From this perspective, neither the Earth nor the other terrestrial planets became giant planets because they did not achieve masses large enough to approach the critical value before the gas of the solar nebula was lost to either the Sun or interstellar space.

The similarity in the ice and rock masses of the giant planets may be attributed to the insensitivity of the critical core mass to their locations in the solar nebula and its weak dependence on the planetesimal accretion rate. However, a true gas runaway accretion phase appears to have been achieved only for Jupiter and Saturn, whose present gas masses exceed those of their ice and rock components. Conversely, Uranus and Neptune appear not to have quite reached critical core masses before the solar nebula was dissipated.

After Jupiter and Saturn achieved a critical core mass and runaway gas accretion commenced, they very rapidly added gas from the surrounding solar nebula. The amount of gas they ultimately accreted may have been limited by several processes. First, once the nearby gas was exhausted, the rate of gas accretion may have been limited by the time needed to transport gas from more distant places in the solar nebula to nearby places. Second, once the planet's mass became large enough, it may have exerted a strong enough gravitational tug on the surrounding nebula to push the nebula away from itself. This non-intuitive repulsive gravitational torque arises in multi-particle media, where frictional forces among the particles exist. This process is responsible for the ability of satellites to create gaps in the rings of Saturn.

At the time that the giant planets finished growing, they were much bigger than their current dimensions. The combination of their self-gravity and their cooling by radiation to space led them to commence a contraction phase that has continued to the present. At first this contraction was very rapid, but it became progressively slower as a larger fraction of the envelope attained high densities at which they started to behave more like an incompressible liquid than a compressible gas. During an intermediate stage in the contraction phase, a flattened disk of gas and dust, derived from either the solar nebula or their outer envelopes, developed around them. Satellites and rings formed in these circumplanetary nebulae. The above phases of growth and contraction in the history of the giant planets are summarized in figure 5-5.

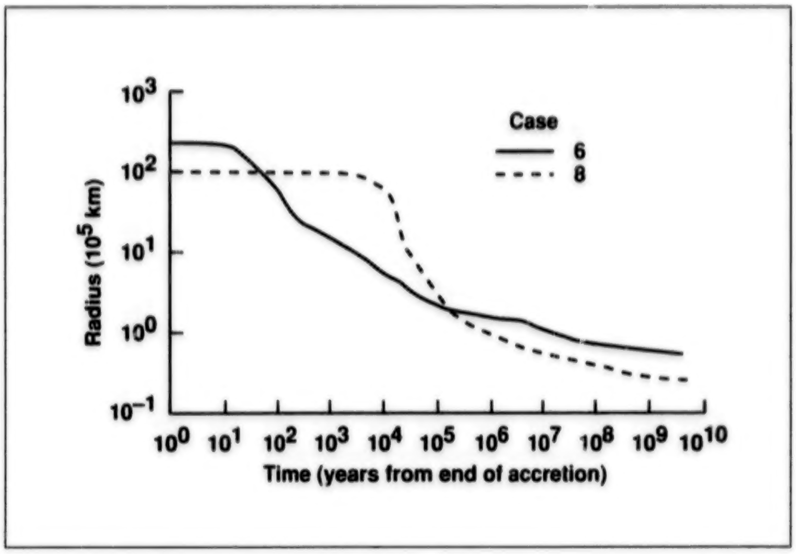

Figure 5-5. Radius of Saturn (solid) and Uranus (dashed) as a function of time from the completion of their accretion.

Composition of the Solar Nebula

The abundance of elements in all parts of the solar nebula is thought to have been identical to that in the early Sun and therefore close to that in the present Sun (only nuclear burning has altered the abundance of a few elements in the Sun). However, the abundances of elements incorporated into the planets that formed within the solar nebula depended on the partitioning of these elements among different chemical species in the solar nebula, the phase of these compounds (gas versus solid), and the mode of the planets' formation.

Of particular interest from an exobiological perspective is the partitioning of carbon, oxygen, and nitrogen atoms among various species and phases in the solar nebula. In strict thermodynamic equilibrium, carbon would have been sequestered almost entirely into carbon monoxide in the warmer regions of the inner solar nebula and into methane in the outer part of the nebula. Temperatures within the regions of the solar nebula where planets formed were probably too elevated for either of these gases to have condensed. However, there are reasons for suspecting that significant departures from thermodynamic equilibrium may have occurred, especially in the colder parts of the nebula. First, chemical reactions may have occurred too slowly in the outer part of the solar nebula, even over its lifetime of several million years, to convert carbon from one chemical form (specifically CO) into its thermodynamic equilibrium form, methane. In particular, carbon monoxide flowing into the outer part of the solar nebula from either the inner part or from the molecular cloud from which the solar nebula formed may simply have remained in the form of carbon monoxide.

Second, complex organic molecules have been found in significant abundances in both certain types of meteorites, the carbonaceous chondrites, that are derived from the outer part of the asteroid belt, and in comets, such as comet Halley. This finding suggests that some carbon in the solar nebula was present as solid organic matter. It was not made biologically, but rather through a series of chemical transformations that occurred in interstellar space (and hence was present in the molecular cloud that gave birth to the solar nebula) and/or the solar nebula. Examples of such processes include ion-molecule reactions that produce observed complicated carbon molecules in interstellar space and the Fischer-Tropsch reaction, whereby carbon monoxide and hydrogen produce organic molecules in the presence of a solid catalyst.

In a similar vein, elemental nitrogen would be expected to have been in the form of molecular nitrogen in the inner solar nebula and ammonia in the outer solar nebula, if thermodynamic equilibrium was the controlling factor. However, slow reaction kinetics might have prevented ammonia from being the dominant nitrogen-containing species in the outer solar nebula. Also, some nitrogen may have been incorporated into organic molecules. However, it seems that nitrogen was less efficiently emplaced into organic compounds in the early solar system, based on the elemental composition of organics found in carbonaceous chondrites and comets.

Almost all the oxygen in the solar system was divided between water, carbon monoxide, and rock. Since the solar elemental abundance of carbon is about half that of oxygen, the abundance of water in the solar nebula depended strongly on the partitioning of carbon into its various possible forms. If no carbon was in the form of carbon monoxide, the water-to-rock ratio in the solar nebula would have been about 1.5, whereas if carbon was entirely in the form of carbon monoxide this ratio would have been about 0.4. In the inner solar system, almost all of the water would have been in the vapor phase, although some hydrated solid silicates could have been produced in the lower temperature portions of this region. In the outer parts of the solar nebula, where temperatures were less than half the room temperature value, water would have been almost entirely in its solid condensed state. Thus, the planetesimals forming the terrestrial planets contained little, if any, water, whereas those that formed the giant planets were rich in water.

Constraints on the partitioning of carbon among various species in the solar nebula can be derived from the composition of current solar system objects. We have already mentioned that the carbon-to-hydrogen ratio in the atmospheres of the giant planets exceeds the solar ratio by factors of few to several tens and that this enrichment of carbon is due to the partial mixing of planetesimal carbon into the envelopes of these planets. By modeling the observed carbon to hydrogen ratio, it is possible to estimate the fraction of the carbon in the outer solar nebula that was contained in the solid phase, perhaps mostly in the form of organics.

During the early growth of the giant planets, when their core masses were less than a few Earth masses, their envelopes were not very massive and all but the tiniest planetesimals would have been able to penetrate through the envelope and reach their cores relatively intact. However, during the later growth stages, when their core masses began approaching their critical values, all but the very largest planetesimals would have been vaporized in their now massive, hot envelopes. Over the entire course of their formation, perhaps 50 to 75% of all planetesimals would

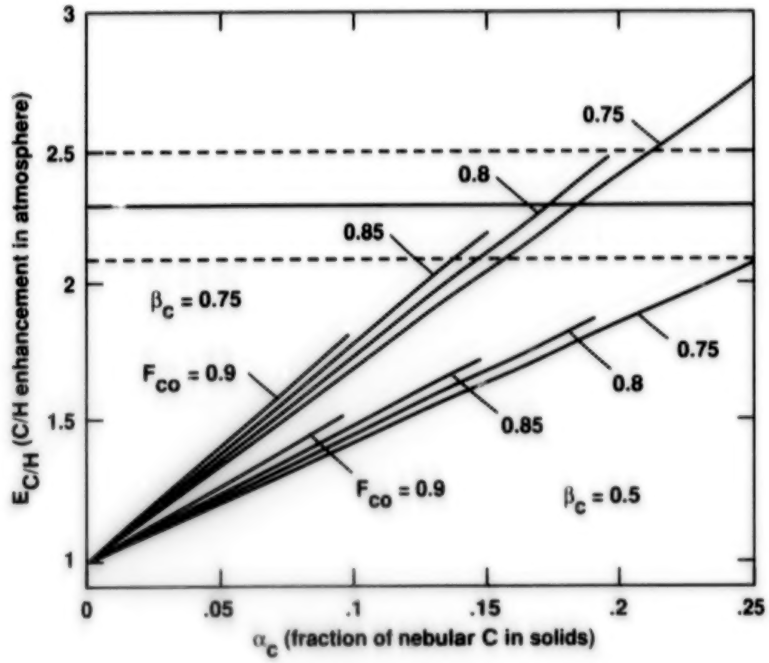

Figure 5-6. Predicted C/H ratio in the atmosphere of Jupiter (in units of the solar ratio) as a function of the fraction of the carbon of the outer solar nebula that was contained in solid bodies. The solid horizontal line shows the observed ratio and the dashed lines alternative values allowed by measurement uncertainties. The various slanting lines show predicted values for various choices of the fraction of nebular carbon in the form of CO, F_{CO}, and the fraction of planetesimals that dissolved in Jupiter's envelope during its formation, β_C.

have vaporized in their envelopes, thus providing carbon for their present atmospheres.

The fraction of nebular carbon, α_c, that was contained in giant planet-forming planetesimals can be found from the observed atmospheric abundance of carbon, the fraction of mass contained in the rock and ice components for the entire planet, and a reasonable value for the fraction of planetesimals that were dissolved in the planet's envelope during its formation, β_c. Figure 5-6 illustrates the value of α_c derived from data on Jupiter when β_c is set equal to 0.5. In this case, α_c equals about 0.2. Comparable values hold for the zones where the other giant planets formed, based once again on their observed atmospheric carbon abundances.

Estimates of the oxidation state of the gaseous carbon species in the outer solar nebula, i.e., the relative abundances of carbon monoxide and methane, may be obtained from the mean density of the solid planet Pluto. By virtue of a relatively unique orbit, Pluto crosses the orbital distance of Neptune, but never comes close to it. Thus, it may be one of the largest surviving planetesimals from the epoch of the formation of the giant planets. Pluto is also unique in having a close moon, Charon. In the last several years, Pluto and Charon have passed in front of one another, as viewed from the Earth. By measuring the characteristics of these mutual occultation events, astronomers have been able to derive a mean density for the Pluto/ Charon system, which is approximately the same as that of the more massive Pluto. Pluto's mean density is about 2 grams per cubic centimeter or about twice that of liquid water (at 4°C).

The mean density of a solid body in the outer solar system is determined primarily by the relative proportion of water and rock that it contains. If it was composed solely of water, its mean density would be about 1 gram per cubic centimeter, whereas if it were made entirely of rock, this value would be about 3.7 grams per cubic centimeter. Pluto's observed density implies that its bulk water and rock fractions are about 0.3 and 0.7, respectively.

As mentioned earlier, oxygen was partitioned in the solar nebula chiefly between water, carbon monoxide, and rock, with the implied relative abundance of water and rock varying significantly as the carbon monoxide abundance varied within limits allowed by solar elemental abundances. The water and rock fractions of Pluto inferred from its mean density indicate that much of the carbon in the region where it formed was in the form of carbon monoxide, as illustrated in figure 5-7.

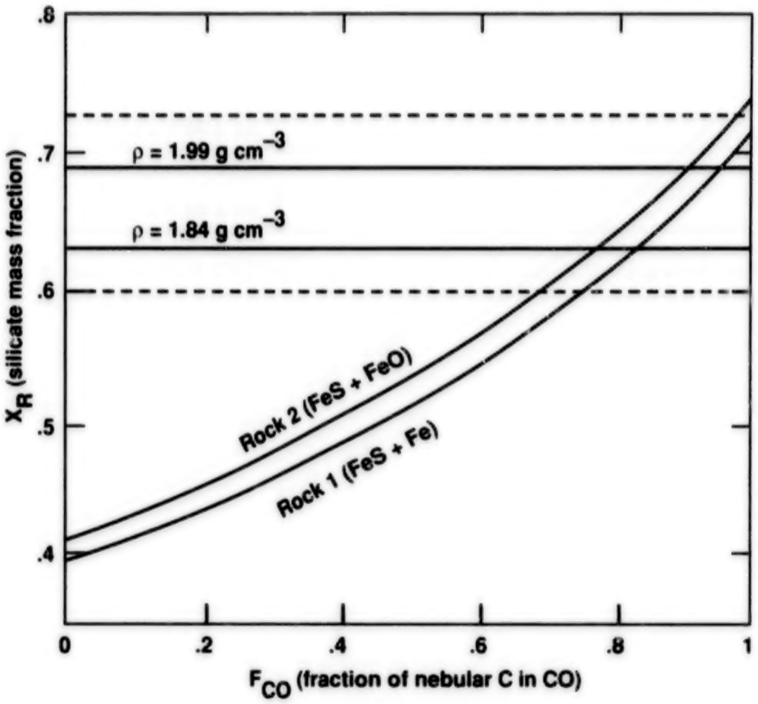

Figure 5-7. The fractional abundance of the rock component of Pluto (the rest is water) as a function of the fraction of carbon that was in the form of gaseous CO in the outer solar nebula. Two possible values of the rock fraction, as derived from Pluto's measured mean density, ρ, are shown by the solid horizontal lines, with their associated uncertainties (dashed horizontal lines). The two slanted lines show the predicted rock fraction for two choices of the composition of the rock.

The above discussion indicates that gaseous carbon monoxide was the chief carbon-containing species throughout the solar nebula. In the region where the giant planets formed a smaller fraction of carbon was present in the solid phase, perhaps chiefly as organics. By implication, gaseous molecular nitrogen was the chief nitrogen-containing species throughout the solar nebula, with a small fraction of the nitrogen (but less than the carbon fraction) in the solid phase.

Implications for Earth

Compared to solar elemental abundances, the Earth as a whole is highly deficient in the materials that are essential to life: water, carbon-containing, and nitrogen-containing species. The ratio of these compounds to rock for the Earth as a whole is about 0.00001 to 0.0001 by mass, as compared to values of about 0.1 to 1 from solar abundances. Yet, there is enough of these materials, particularly since they are concentrated near the Earth's surface, to have permitted life to have arisen some 4 to 4.5 billion years ago and to have sustained it until the present. By comparison, the giant planets and Pluto are rich in these volatile compounds, having abundances relative to rock that are comparable to that expected from solar abundances. Yet, no life has been detected on them to date.

The above stark difference in volatile abundances between the inner and outer solar system raises the question as to the ultimate source region for the volatiles that the Earth presently possesses. On the one hand, only very modest amounts of volatiles in the planetesimals that formed near the Earth and contributed the bulk of its mass would have sufficed to account for their present abundances. On the other hand, relatively small contributions to the Earth's mass from volatile-rich bodies formed farther out in the solar nebula could have provided the bulk of the planet's volatiles. For the moment, let us suppose that the latter is true and consider the role that the giant planets played in transferring volatile-rich planetesimals from the outer to the inner parts of the solar system.

When small bodies pass close to, but do not collide with, a planet, they have their direction of travel altered by the gravitational field of the planet. For a planet to be able to transfer the orbit of a planetesimal from its region of the solar system to that of other planets, the escape velocity from its surface (a measure of the strength of its gravitational field) needs to be comparable to or greater than the orbital velocity of the planetesimal. The giant planets met this criterion once they attained their current masses and once they contracted to a size comparable to their current size.

More precisely, Uranus and Neptune were then able to scatter planetesimals from their neighborhood either into the more distant regions of the solar system, where they may have formed the Oort cloud from which comets are derived, or to the orbit of Saturn. Saturn and, particularly, Jupiter were able to scatter the once scattered planetesimals originating near Uranus and Neptune, as well as their own planetesimals, into the inner part of the solar system. These scattered planetesimals, as well as ones removed from the asteroid belt directly or indirectly by Jupiter, were the principal contributors to a period of heavy bombardment that the terrestrial planets experienced during their first 700 million years of existence. During this epoch the terrestrial planets and the Moon were cratered by impacting bodies at a rate that was at least a thousand times greater than the current rate.

Thus, over a period that extended from the end of the Earth's formation through its early history, the Earth encountered a large number of bodies, some of which may have been very volatile-rich and therefore may have contributed significantly to its present volatile inventory. These bodies originated in or near the region of the outer planets and were scattered by them into the inner solar system. In this plausible, but not yet proven scenario, living organisms on Earth today, including ourselves, may be made in part of atoms that were situated at one time close to the region of the outer planets and that were brought to the Earth through the gravitational effects of these planets.

Additional Reading

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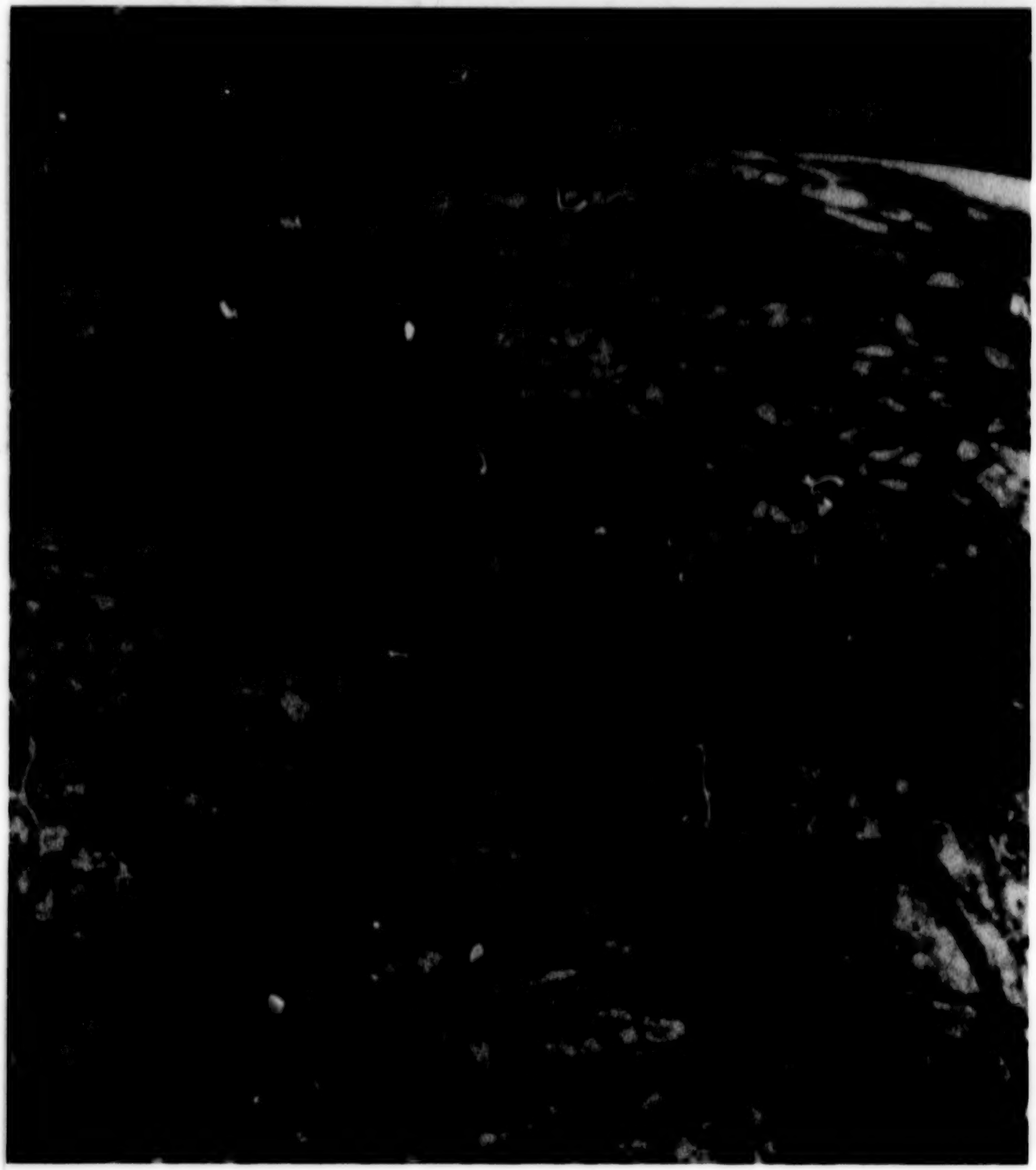

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Europa: Prospects for an Ocean and Exobiological Implications

N93-18552

J. Oro, S. W. Squyres, R. T. Reynolds, and T. M. Mills

Introduction

As far as we know, Earth is the only planet in our solar system that supports life. It is natural, therefore, that our understanding of life as a planetary phenomenon is based upon Earth-like planets. For example, Mars has received a great deal of attention as a possible former abode for life because its climate appears to have been somewhat more Earth-like in the past. However, there are environments in the solar system where liquid water, commonly believed to be a prerequisite for biological activity, may exist in a distinctly non-Earth-like environment. One such location is Europa, one of the Galilean satellites of Jupiter. The possibility that liquid water exists on Europa presents us with some interesting exobiological implications concerning the potential of the satellite to support life.

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An Ocean on Europa?

The images of Europa taken during the Voyager flybys of the Jovian system (chapter frontispiece) show a very bright surface transected by a network of long linear features of lower albedo and brownish color. The features are seen down to the resolution of the images (~4 km/line pair) and are as large as tens of kilometers across with maximum lengths comparable to the radius of Europa (~1500 km). Their geometry indicates that they are probably fractures caused by extensional stresses in Europa's crust. Other linear topographical features include low ridges of unknown origin. Overall, however, the surface of Europa is remarkably level. Only a handful of small impact craters have been positively identified, and it is doubtful that topography anywhere on the satellite exceeds a few hundred meters.

Even the most basic pre-Voyager information about Europa appeared enigmatic. The surface is composed primarily of H_2O and frost, yet the satellite's density clearly indicates that it is predominantly a silicate body. Simple but powerful cosmochemical arguments suggest that all of the Galilean satellites should be composed primarily of silicates and varying amounts of H_2O. Water frost was first positively identified as a major component of Europa's surface in 1972, based on the presence of strong infrared H_2O absorption features in its reflectance spectrum. This finding has been confirmed by later observations, which have led to the conclusion that more than 95% of the spectroscopically detectable material on the surface of Europa is water. Despite the nearly pure water ice surface composition, the density of Europa is known to be 2.97 g cm^{-3}. A density this large indicates a composition dominated by silicates, with only a relatively small admixture of H_2O. If the silicate component of Europa is largely dehydrated and has a density like that of Io (3.57 g cm^{-3}), then Europa is composed of 7 to 8% free H_2O by mass. If, however, the silicates are hydrated, then the density is consistent with a composition including little or no free H_2O.

Three models have been proposed for the internal structure of Europa that are consistent with the satellite's surface composition and density. These are the *thin ice, thick ice,* and *ice/ocean* models (fig. 6-1). The thin ice model suggests that the silicates in Europa's interior are largely hydrated and that the surface H_2O ice is a very thin layer (a few km) lying over the hydrated silicates. In the thick ice model, enough internal heat is produced and retained to dehydrate the silicates, driving the H_2O to the surface to form a layer of solid ice on the order of 100 km thick. In the ice/ocean model still more heat is produced and retained, melting much of the ice. The internal structure of Europa in this model consists of a dehydrated silicate interior, an ocean of liquid water on the order of 100 km thick, and a thin (≤10 km) ice layer on the surface.

All of these models are consistent with Europa's density and surface composition and choices among them must be made on the basis of surface morphology and models of Europa's internal thermal evolution.

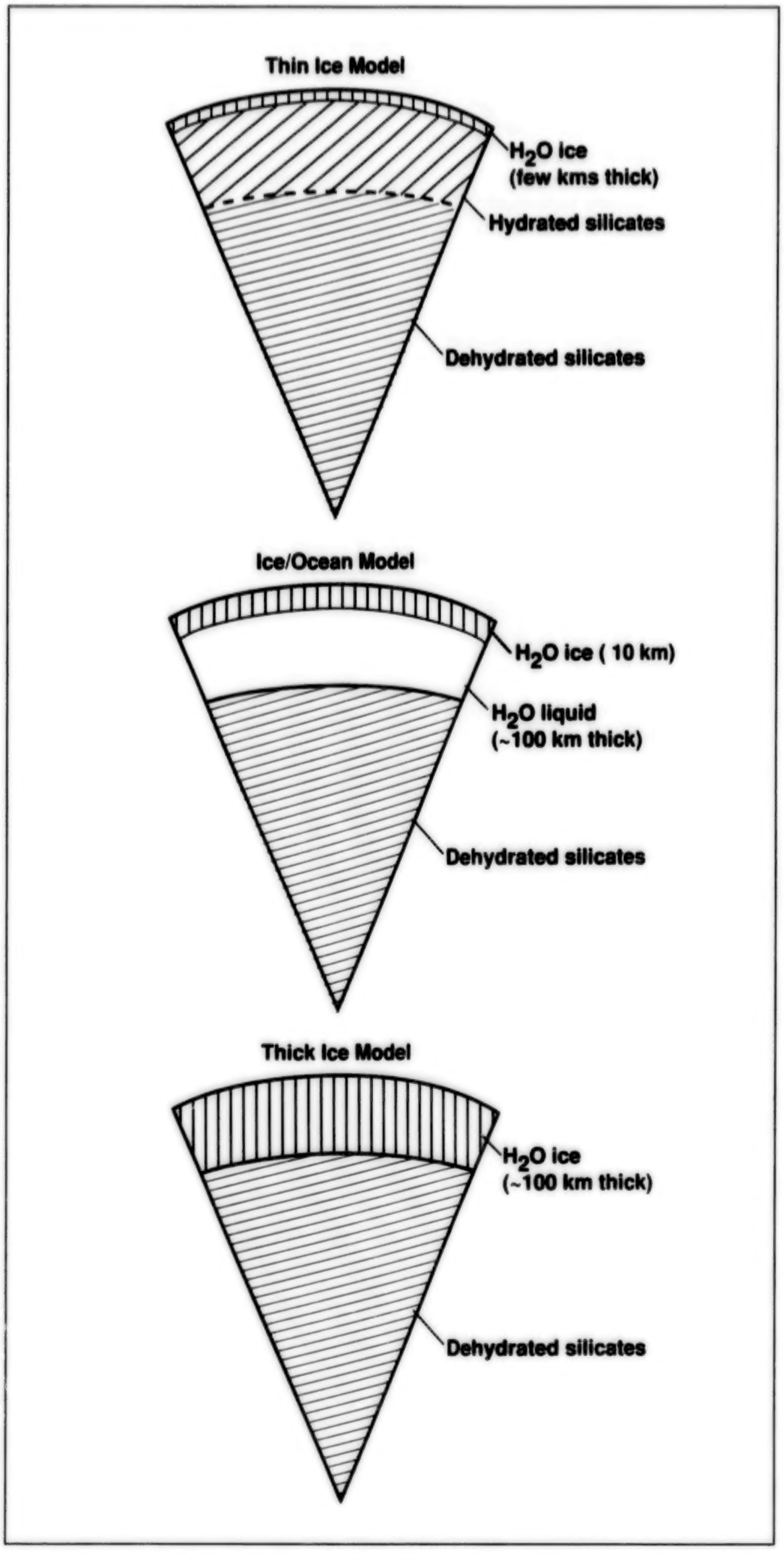
Thin Ice Model
H_2O ice
(few kms thick)
Hydrated silicates
Dehydrated silicates
Ice/Ocean Model
H_2O ice (10 km)
H_2O liquid
(~100 km thick)
Dehydrated silicates
Thick Ice Model
H_2O ice
(~100 km thick)
Dehydrated silicates

Figure 6-1. Three models proposed for the internal structure of Europa include the thin ice, thick ice, and ice/ocean models.

Thermal Evolution of Europa

The heat source that may maintain liquid water within Europa is the dissipation of tidal energy. Tidal forces arise in a satellite because the gravitational field of the planet varies across the body of the satellite. Regions on the satellite that are nearer the planet feel a stronger attractive force than those that are farther from the planet. The effect of these forces is to distort the satellite's shape. Tidal torques generated by this distortion rapidly fix the satellite's rotation to be synchronous, so that one face always points in the direction of the planet. However, if the orbit is eccentric, the size of the tidal bulge varies with distance from the planet. Hence the satellite is continuously flexed as it proceeds around its orbit. Conservation of angular momentum and energy in this process results in the eccentricity of the orbit being damped, so that the orbit gradually becomes circular unless some process acts to maintain the eccentricity. Hence the tidal heating process is self-limiting and the heating will cease when the circular orbit condition is reached.

However, for many of the satellites of the outer solar system, mutual interactions among the satellites act to maintain non-circular orbits over extended periods. The prime example of this phenomenon is the "Laplace resonance" between Io, Europa, and Ganymede. This resonance results in the continuous forcing of Europa's eccentricity, and thus, the continuous deposition of tidal energy as heat. Physically, the energy dissipated by the tides comes from the orbital energy of the satellite and the rotational energy of the planet.

The mean heat flow for Jupiter's innermost large satellite, Io, has been determined observationally to be ~1500 erg cm^{-2} s^{-1}. This heat flow is driven by tidal dissipation and is reflected in the spectacular volcanic activity observed on the surface of Io by Voyager. However, the tidal dissipation rate is also strongly dependent on the orbit's semi-major axis, a, which is defined as half the length of the longest axis of the orbital ellipse. The tidal dissipation rate decreases as $a^{-15/2}$. So Europa, largely because it is significantly farther from Jupiter than Io, undergoes substantially weaker tidal heating. The crucial question is whether or not the heating is sufficient to maintain liquid water in the satellite's interior.

Europa's Three Models

Consider again the three possible models of Europa's interior discussed above (fig. 6-1). Of these, the thin ice model appears to have the most severe problems. Radiogenic heat is generated by a body as a result of the decay of radioactive isotopes trapped in the interior of the planet or satellite upon formation. When one calculates the magnitude of tidal heating in the silicate portion of Europa it appears that the combined tidal and radiogenic heat flow is sufficient to dehydrate the bulk of the silicates in Europa's interior, thus producing an amount of water or ice at the satellite's surface in excess of that predicted by the thin ice model. More importantly, it has been shown that the thin ice model should lead to long-term retention of large impact craters since rock close beneath the surface would inhibit glacier-like relaxation topography. This result is in strong contrast with Europa's nearly crater-free surface. Choosing between the thick ice and ice/ocean models, however, is more difficult. The critical problem concerns the rate of heat removal by thermal conduction and solid state convection in the ice. According to tidal heating calculations, conduction alone cannot remove heat rapidly enough from Europa to keep most of the ice in a thick outer layer from melting. On the other hand, if solid state convection occurs through the ice, it can remove heat rapidly enough to prevent melting and maintain a solid ice layer. On Earth, flow of glaciers takes place by deformation of ice in the solid state; similar deformation could allow slow convection in Europa's icy crust. However, solid state convection in an outer ice layer can occur only if the layer exceeds some critical thickness, which is about 30 km on Europa.

Therefore, the crux of the dilemma in choosing between the two models lies in the question of whether the tidally generated heat flow ever became low enough to allow the freezing of Europa's entire water mass. If this occurred then Europa has probably lost its heat effectively through time by solid state convection in the ice, avoiding the formation of an ocean. Alternatively, if the heat flow never became small enough to allow the formation of a layer of ice 30 km thick then an ocean should exist. The period of Europa's thermal evolution in which this potential evolutionary bifurcation is believed to have been resolved is the period toward the end of the satellite's accretion. However the outcome of a particular evolutionary model that predicts the conditions of the satellite toward the end of this critical accretion period hinges on the highly uncertain calculation of the tidal heating rate. Very simple calculations suggest that tidal heating rates exceed those that would allow convection and freezing by more than a factor of two. More detailed consideration of the rheology of an ice shell has suggested that tidal heating in such a shell would be smaller than this, and preservation of a liquid layer correspondingly more difficult. On the other hand, consideration of tidal dissipation in an ice layer of varying thickness has led to the suggestion that these variations could lead to enhanced local heating and melting. Based on theoretical arguments alone, then, the presence of a liquid ocean appears plausible but not proven.

If an ocean does exist, fractures that occurred through the entire depth of the ice layer would briefly expose the liquid water below. This possibility is supported by the existence of apparent fractures on the surface of Europa. Fracturing could be caused by tidal stresses or by movement of the shell with respect to the tidal bulge of Europa. Liquid water exposed to the surface by fracturing would boil rapidly, creating a cloud of vapor that would condense and fall as frost over the surface of the satellite (fig. 6-2). An insulating layer of porous frost would reduce the surface conductivity below that of solid ice, thereby lessening the escape of heat which would lead to increased melting and the further reduction of the thickness of the ice shell.

Several lines of observational evidence tentatively support the hypothesis of liquid water below a thin ice shell on Europa as well as active resurfacing of the surface, possibly by release of water through cracks in the ice. First, the lack of craters and the apparent mobility of the crust are most readily explained in this manner. Second, the photometric function of Europa's surface indicates that there is a substantial textural difference between that surface and impact ejecta deposits of similar albedo on Ganymede and Callisto. The photometric properties may be more consistent with a surface layer of condensed frost as opposed to impact ejecta. Third, the SO_2 concentration observed on Europa's surface can be interpreted to be a result of uniform addition of S^- ions from Jupiter's magnetosphere to the surface on which a much larger amount of H_2O is continually deposited. Deposition rates for H_2O of $0.1\ \mu m\ yr^{-1}$ are inferred. Despite these inferences, however, the hypothesis that an ocean of liquid water exists beneath Europa's icy shell remains unproven. Definite conclusions must await further exploration of the satellite.

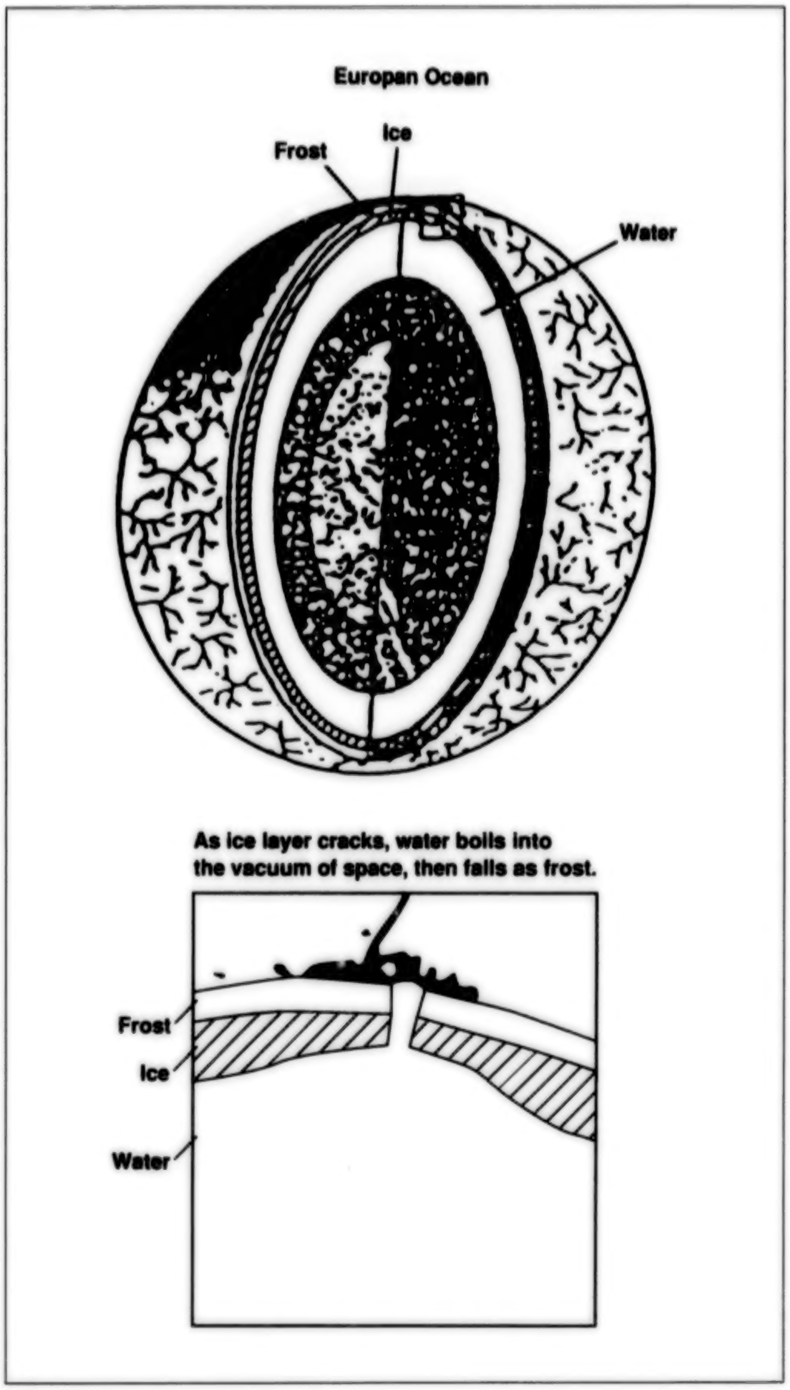
Europan Ocean
Frost
Ice
Water
As ice layer cracks, water boils into the vacuum of space, then falls as frost.
Frost
Ice
Water

Figure 6-2. Liquid water exposed to the surface by fracturing would boil rapidly, creating a cloud of vapor that would condense and fall as frost over the surface of the satellite.

Exobiological Implications

If an ocean does exist in Europa, what would be the exobiological implications of this fact? We know that in addition to the presence of liquid water and biologically useful sources of energy, the presence of compounds of the biogenic elements (H, C, N, O, S, P) is considered essential for the existence of life in any cosmic body.

Very little is known about the bulk composition of Europa and other Galilean satellites except that they must have been made of matter present in the Jovian nebula, rich in water and other compounds of the biogenic elements, as suggested by studies on the formation and evolution of these satellites. It has been argued that as a result of the substantial luminosity of proto-Jupiter the condensation of ices, such as H_2O, was inhibited at a close distance to the planet, and a sharp chemical gradient was created among the Galilean satellites. Under these circumstances Io would only have retained dehydrated silicates and other minerals, sulfur, and SO_2, whereas increasing amounts of H_2O in the form of ice or water of hydration would have been retained by Europa, Ganymede, and Callisto.

Table 6-1: Density Comparisons between Galilean Satellites and Carbonaceous Chondrites*

Galilean satellites		
Io	Europa	Ganymede
3.57	2.97	1.93
Carbonaceous chondrites		
Type III	Type II	Type I
3.5	2.5-2.9	~2.0

*Density comparisons between carbonaceous chondrites and the Galilean satellites ignore the processes of self compression and ice phase changes that presumably took place on the satellites.

The only direct evidence available about the composition of these satellites is that obtained by spectral measurements on their surfaces. Io's surface is dominated by sulfur and its compounds, particularly condensed sulfur dioxide. The other three satellites have copious amounts of water-ice in their surfaces, about 95% in Europa, 90% in Ganymede and 30 to 90% in Callisto. The non-water ice material on the Galilean satellites appears to be spectroscopically similar to that of carbonaceous chondrite material and to that of minerals containing Fe^{3+}. Carbonaceous chondrites (particularly Wiik Types I and II) are a class of meteorites characterized by their high concentrations of iron-magnesium silicates and of volatile components such as water, sulfur, and carbon in comparison to other meteorite classes.

On the basis of these and other observations a suggestion is advanced, as a working hypothesis, that the Galilean satellites were made of matter similar to that of carbonaceous chondrites but with increased amounts of water and carbon compounds as would be expected from the composition of the Jovian nebula. The suggestion is also consistent with the relative proximity of the Jovian nebula to the outer region of the asteroid belt where most of the dark asteroids exist. Dark asteroids are presumed to be the parent bodies of the carbonaceous chondrites and, in fact, if we compare the Galilean satellites with carbonaceous chondrites, we see an approximate correlation of their densities. The densities of Io (3.57 gm cm^{-3}), Europa (2.97 gm cm^{-3}), and Ganymede (1.93 gm cm^{-3}) are very similar to the densities of Type III (3.5 gm cm^{-3}), II (2.5-2.9 gm cm^{-3}) and I (~2.0 gm cm^{-3}) carbonaceous chondrites, respectively, according to Wiik's classification (table 6-1). The approximate correlation of the densities is appropriate and useful in that it allows the estimation of the composition of Europa using available data. However, it should be pointed out that a direct density comparison between carbonaceous chondrites and the Galilean satellites ignores the processes of self compression and ice phase changes, processes that presumably led to an increase in the densities of the more massive satellites of Jupiter relative to the less massive carbonaceous chondrites, and thus, should be taken only as a demonstration of an approximate relationship.

Table 6-2: Chemical Composition of Carbonaceous Chondrites

	SiO_2	MgO	C	H_2O	S
Type I	22.56	15.21	3.54	20.08	6.20
Type II	27.57	19.18	2.46	13.35	3.25
Type III	33.58	23.75	0.46	0.99	2.27

If we limit our discussion to the case of Europa and compare it with Type II carbonaceous chondrites (CII), we see that in addition to sulfur (3.25%) the CII chondrites contain relatively large amounts of water (13.35%) and carbon compounds (2.46%) (table 6-2). This shows that planetesimals of this composition, which are believed to have been involved in the formation of Europa through collisions with the proto-satellite throughout its early history, would have carried with them more than sufficient amounts of H_2O to have provided the calculated water fraction of Europa which is at least 7% of the mass of the satellite ($M_e = 4.87 \times 10^{25}$ g). But what is more important from an exobiological point of view is that a significant presence of organic compounds in Europa is also suggested by this comparison. What are the natures of the organic compounds? As indicated by the studies done on the Murchison meteorite, one of the best analyzed CII chondrites, a large variety of organic and other carbon compounds are present (table 6-3). The organic matter present is distributed among different compound classes of obvious biological interest, including amino acids, purines, and pyrimidines, which are the building blocks of proteins and nucleic acids, respectively. It should be noted that the relative amount of the soluble organic compound fraction (about 500 ppm in Murchison) would be equivalent to about 0.05% of the mass of the satellite.

Table 6-3: Distribution of Carbon in the Murchison CM2 Meteorite

Species	Abundance
Acid insoluble carbonaceous phase	1.3-1.8%
Carbonate and CO_2	0.1-0.5%
Hydrocarbons	
aliphatic	12-35 ppm
aromatic	15-28 ppm
Acids	
monocarboxylic (C_2-C_8)	~170 ppm
dicarboxylic (C_2-C_9)	not measured
hydroxy (C_2-C_5)	~6 ppm
Amino Acids	10-20 ppm
Alcohols (C_1-C_4)	~6 ppm
Aldehydes (C_2-C_4)	~6 ppm
Ketones (C_3-C_5)	~10 ppm
Ureas	~20 ppm
Amines (C_1-C_4)	~2 ppm
N-heterocycles	
pyridines and quinolines	0.04-0.4 ppm
purines	~1 ppm
pyrimidines	~0.05 ppm
poly-pyrroles	<<1 ppm
Sum:	1.43-2.35%
Total carbon:	2.0-2.5%

We can translate the CII meteorite data into a possible composition of the upper hydrous layers of Europa. Assuming that about 90% of the satellite's water is in the liquid phase (namely 6.3% of the mass of the satellite) the concentration of organic compounds dissolved in such a subsurface ocean would be of the order of 1%. In absolute terms the amounts of total carbon (carbonates plus carbon compounds) in Europa would be 1.2×10^{24} g and that of soluble organic compounds 2.4×10^{22} g. The latter figure is at least four orders of magnitude higher than the total organics in Earth's active biosphere. Moreover, a concentration of 1% organics in water is a sufficient concentration to lead to significant reaction rates for abiotic biochemical synthesis under favorable conditions. Whether these reactions could have led to emergence and evolution of life on Europa is not known and would have depended very much on the conditions of the satellite toward the end of the accretion phase, as discussed earlier.

Considering what may be occurring today on the surface of Europa, possible evidence for the synthesis of organics may be visible on the satellite's surface. The presence of brown colorations on the surface fractures of Europa has led to the speculative suggestion that the colorations present are due to organics produced abiotically on or below the surface of Europa. Organics synthesized below the surface could be brought to the surface as the result of fractures in the ice. However, at this time the results from a number of investigations that in some way model the conditions on Europa have produced only highly speculative implications.

Early Conditions of Europa

What were the early conditions of Europa and how do they compare with those of carbonaceous chondrite parent bodies? The occurrence of an aqueous environment on such a parent body has been proposed before based primarily on the general characteristics of carbonaceous chondrites. A similar model has been used to explain the similarity of the amino acids found in the Murchison meteorite to those obtained from the classic Miller experiments with electrical discharges. Thus it has been suggested that a parent body of about 100 km in radius could have retained a liquid environment for about 200 million years and an atmosphere for at least 10^4 years. If that were the case, such a parent body model would explain the observations mentioned above on the similarities between amino acid compositions of carbonaceous chondrites and those produced under laboratory models of the prebiotic Earth. These environmental conditions would have been much more plausible for Europa because of the much larger mass and radius (1,563 km) of this satellite as well as its relative proximity to Jupiter.

Studies on the evolution of the Jovian system and the Galilean satellites suggest that during its early phase of formation, Europa was immersed in the Jovian nebula for 10^5 to 10^6 years and its atmosphere was probably retained for a significantly longer time. Therefore, during this time, most of Europa's water was probably partially solid and liquid, coexisting with its atmosphere, an atmosphere which should have been reducing in nature and similar to that of Jupiter. Thus, a number of experiments performed in different laboratories on the abiotic synthesis of organic compounds may be relevant to the early Europan conditions. These studies include the work done by several investigators with electrical discharges and different sources of radiation. Interestingly, these experiments are similar to those described earlier which were performed to explain the current existence of colored compounds along the linear markings of Europa.

Additional synthetic chemical reactions could have occurred upon the infall of planetesimals, meteorites, and comets on Europa. The organic compounds present in the satellite or in the infalling bodies would have been subjected to heating and rapid quenching and cooling at the water-surface interface upon collision during the last phase of satellite differentiation. The low temperatures of Europa's hydrous phases would have facilitated the survival and further interaction of the biochemical compounds which were generated in this early stage.

Low-Temperature Abiotic Chemistry

Examples of relevant low-temperature, abiotic reactions are offered by the synthesis of adenine and purines from hydrogen cyanide and by the formation of peptides by condensation of amino acid amides. Hydrogen cyanide is a ubiquitous cosmic precursor of biochemical compounds since it is present in many cosmic bodies, including interstellar clouds, comets, Titan, Jupiter, and Saturn, and must have been present in the Jovian nebula as well. The formation of adenine and other purines from hydrogen cyanide is well known from our earlier work. Adenine is probably present in Halley's comet as shown by data from the spacecraft Vega 1 (U.S.S.R.) mass spectrometer. This important biochemical compound has also been speculated to be present at the bottom of Titan's methane-ethane ocean.

What is less well known is that hydrogen cyanide can produce the precursors of purines in high yields at temperatures of −10°C and lower from dilute solutions (0.001 M) of HCN. Upon further cooling and formation of the water-ice, the hydrogen cyanide solution is concentrated until it becomes 75% HCN at the eutectic point (−22°C). Thus by freezing, enormous enrichments are possible of reactive chemical species which produce biochemical monomers. A similar situation applies to the condensation of monomers to polymers observed in the conversion of glycinamide, a common precursor of glycine, to polyglycine, a linear molecule containing a number of amino acid residues joined end to end. Glycinamide is also readily obtained from hydrogen cyanide. Furthermore, the model proposed for the replication of polynucleotides also requires relatively low temperatures (0 to 25°C). In fact, a number of investigators consider the low-temperature prebiotic synthesis of biochemical compounds a crucial stage for the origin of life on Earth.

Possibility of the Emergence of Life on Europa

As discussed above, the early history of this Galilean satellite may have been favorable for the capture and *in situ* formation of different organic compounds, including biochemical monomers and polymers. The low-temperature environmental conditions may have also facilitated the preservation and further interaction of labile biochemical polymers.

Whether these processes proceeded to a more advanced stage of development is not known because no experimental work has been done at low temperatures on the protocellular stages of prebiological evolution. A possible difference with the conditions of the primitive Earth may be the absence of processes of cyclic evaporation for a continued period of time under atmospheric pressure. This is hypothesized to have facilitated on our planet the formation of phospholipids and other amphiphilic compounds which presumably gave rise to liposomes that encapsulated catalytic nucleotides. Whether these and other stages of protocellular evolution may or may not have been possible on Europa is not clear. However, recent information obtained on the amphiphilic components of the Murchison meteorite has shown that they have the ability to self-assemble into bilaminar and trilaminar membranes similar in structure to the membranes found in living cells. This is a significant finding that increases the possibilities of formation of membranous structures which could facilitate the appearance of life on Europa.

Table 6-4: Substrates and Energy Available for Metabolic Activity

A. Substrates	Wt %*	Mass
Total carbon (carbonates + C compounds)	2.46%	1.2×10^{24} g
Total sulfur	3.25%	1.6×10^{24} g
Free sulfur (Mighei)	2.39%†	1.2×10^{24} g
Total phosphate (as P_2O_5)	0.27%	1.3×10^{23} g
Soluble phosphates (~10% of total phosphate)	0.027%	1.3×10^{22} g
Soluble organic compounds	~0.05%	2.4×10^{22} g
Soluble nitrogen compounds (Orgueil, CI)	0.01%	4.9×10^{21} g

B. Energy sources for metabolic activity	
Hot suboceanic thermal springs	>100°C
Chemical energy from redox reactions; thermal plumes with unequilibrated chemical systems	e.g., H_2, H_2S, S, reduced organics, SO_2, SO_4, CO_2
Solar radiation	Surface and upper layers

*Wt % are averages for CII carbonaceous chondrites except for soluble nitrogen (CI).
†Wt % is average from Mighei analyses.

Prerequisites for the Habitability of Europa

Leaving aside the unanswered question concerning the possible origin of life on Europa, it is also of interest to determine whether any form of indigenous or terrestrial life may be able to survive and reproduce under the existing conditions of this satellite. The presence of liquid water is one of the major requirements for the existence of life in a cosmic body. As discussed above, Europa may have possessed a significant body of water since its formation and during the course of its geological history so that this first requirement may be fulfilled. Two other necessary requirements are the availability of compounds of biogenic elements (as precursors of biochemical molecules and as substrates for metabolic reactions) and a continuous supply of energy.

As discussed earlier, we have seen that using the carbonaceous chondrite model, about 2.4×10^{22} g of soluble organic compounds may have been present at the time of formation of Europa (tables 6-3 and 6-4). Furthermore, the total amount of organic and inorganic carbon is about 1.2×10^{24} g and some of this carbon (e.g., CO_2) could, in principle, be available to enter into metabolic cycles. As seen in table 6-4, the total value for sulfur (in any form, reduced and oxidized) would be expected to be of the same order of magnitude as that for total carbon.

On the other hand, the availability of the biogenic element nitrogen is not so well known. Based on the amounts of amino acids and heterocyclic nitrogenous compounds in the Murchison meteorite, the values for soluble organic nitrogen are about an order of magnitude lower than those for the dissolved carbon. It should be pointed out, however, that some nitrogen was probably retained in the form of ammonium sulfates, sulfides, or other salts which must have been present in the Jovian nebula. Indeed, Jupiter presumably has large amounts of ammonia and ammonium sulfides, and at least one analysis of the Orgueil meteorite, a CI chondrite, gave a value of 0.098% for ammonium chloride. If a similar percentage had been present in Europa it would correspond to about 4.9×10^{21} g for the whole satellite. Therefore, the total amount of soluble nitrogen on early Europa may have been somewhat lower than that of dissolved carbon (2.4×10^{22} g).

The other, extremely important biogenic element is phosphorus, which has been found in the form of PN in interstellar clouds, phosphine in Jupiter, and phosphates in interplanetary dust particles and meteorites. More specifically, phosphorus is also present in the three types of carbonaceous chondrites. The average amount of phosphate measured as P_2O_5 in the CII meteorites is 0.27%. The corresponding total amount for Europa would be 1.3×10^{23} g (table 6-4). This biogenic element exists in the above meteorites primarily as insoluble and partially soluble calcium phosphate minerals. It is also present in the form of phosphides. Upon solubilization of the inorganic phosphates they would be suitable for participation in the abiotic synthesis of key biochemical compounds, such as nucleic acids and important metabolic substrates. Because they are in the form of chloro- or hydroxyl apatite as well as CaMgH phosphates and possibly as sodium phosphates, only a small percent of these phosphates would need to be solubilized (e.g., ~10%) for the amount of the phosphate anion to reach the value of 1.3×10^{22} g which is more than 50% of the amount of dissolved carbon.

In summary, using the carbonaceous chondrite model, the amounts of reactive carbon, nitrogen, and phosphate compounds available as precursors or metabolic sources for the synthesis of biochemical compounds would range from 4.9×10^{21} to 2.4×10^{22} g. Even assuming that only 10% of these compounds reached the Europan ocean, the subsequent amounts of these elements present in the ocean would be in the range of 10^{20} to 10^{21} g, values which are two to three orders of magnitude larger than the total terrestrial biomass (10^{18} g).

Energy Sources for Biosynthesis and Metabolic Activity

Assuming Europa's ocean exists, there are primarily two major energy sources in the aqueous layers of Europa which could be utilized biologically for the synthesis of biochemical compounds and for different metabolic reactions.

The first is the dissipation of tidal energy in the form of heat, which presumably maintains a significant fraction of the water on Europa as liquid. In Io, the dissipation of tidal energy does not occur in a very uniform manner, but rather in discrete zones or spots, as shown by the more than 100 sulfur volcanoes on its surface. At a much lower scale the transfer of heat from the core of Europa to the bottom liquid water layers of its ocean may also occur in localized places. The existence of these non-uniform areas of heat transfer in Europa are at present only speculative, however, assuming they exist, they could be analogs of terrestrial deep oceanic thermal vents. On the

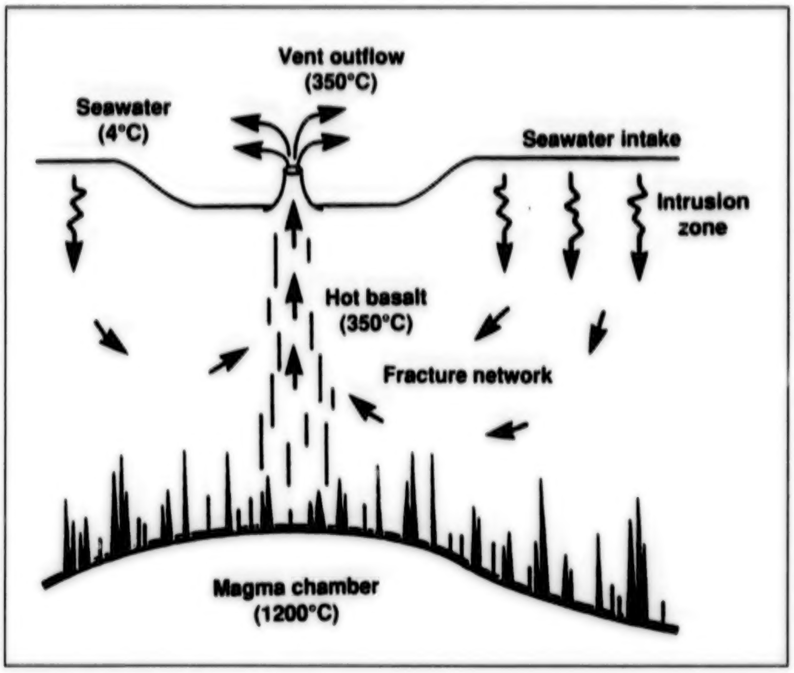

Figure 6-3. On Earth, the concentration of thermal energy in local high-temperature regions at submarine crests has led to large-scale hydrothermal activity. Reactions between sea water and hot basalt beneath the ocean floor produce solutions that ascend to the seafloor and emerge as hot springs whose sulfur and hydrogen compounds sustain chemosynthetic or chemolithotrophic bacteria.

Earth, the concentration of thermal energy in local high-temperature regions at submarine crests has led to large-scale hydrothermal activity (fig. 6-3). Reactions between sea water and hot basalt beneath the ocean floor take place at temperatures in excess of 300°C. The resulting solutions ascend to the sea floor where they emerge as hot springs that sustain oases of life. At the base of the food chain in these oases are chemosynthetic or chemolithotrophic bacteria that derive their entire energy supply from the reaction of geothermally produced sulfur and hydrogen compounds released from the vents.

Analogous hydrothermal regions conceivably could exist on the bottom of Europa's ocean. The calculated mean heat flux from Europa's silicate interior is less than a third of the mean heat flow of the Earth. In fact, it is only some 50% higher than that of the Moon, which is not observed to have any volcanic activity at the present. However, since the Moon's near-surface regions are virtually water-free this does not rule out the possibility of hydrothermal activity on Europa. Current data and modeling techniques are insufficient to provide a quantitative assessment of the probability of such activity.

The existence of sulfur on Europa, as suggested by the carbonaceous chondrite model, offers an additional similarity between terrestrial and theoretical Europan oceanic vents. It is of interest to note in this respect that due to the unequilibrated nature of the matter in these meteorites, and also probably in the Jovian nebula, several of the different states of oxidation of the sulfur compounds may coexist simultaneously, e.g., from the most reduced sulfides to the most oxidized sulfates. This would establish an adequate chemical redox potential ideally suited for utilization in metabolic reactions early in the satellite's development.

The latter infers the plausibility of the existence, at some time during Europa's past or present history, of a microbial ecosystem analogous to those that exist presently near terrestrial submarine geothermal vents.

The second form of energy that could be utilized biologically is the solar radiation which reaches the icy surface of the satellite. Assuming that solar energy would reach liquid water only where the ice crust had recently been fractured, it is straightforward to predict the maximum amount of light that could reach the Europan ocean. Calculations show that the light levels in the water would be extremely low, even immediately after a fracturing event. However, for up to 4.5 years after a fracturing event they could remain higher than the levels at which photosynthesis occurs in primitive microbial mats in perennially ice-covered lakes in the Dry Valleys of Antarctica. So, if such fracturing does indeed occur, very limited local environments may exist for extremely short periods of time that are within the range of adaptation of simple Earth organisms.

Two very important caveats should be added to this discussion of solar energy input. First, the scenario presented here represents a strict upper limit in that it does not consider loss of thermal energy to the walls of a fracture. If a fracture is narrow relative to the thickness of the forming ice cover, heat loss to the walls will cause freezing that may be substantially more rapid than we calculate. Second, we do not know the frequency with which fracturing events occur, or indeed if they take place at all. Using the most liberal estimates of fracturing, based on an inferred resurfacing rate by recondensing frost, we obtain a maximum total fracture area of approximately 5 km^2/yr, which is an exceedingly small fraction of the total surface area of the satellite.

Habitability of Europa by Anaerobic Life

On the basis of discussion so far, the three most important requirements for the maintenance of life (i.e., liquid water, sources of energy and organic compounds) could conceivably have existed on Europa during all of its geological history. Therefore, it may be appropriate to consider ecosystems on Earth that can be used as analogs to a putative Europan biota. Since no significant amounts of molecular oxygen appeared in the Earth's atmosphere until about 2 billion years ago, it would be reasonable to assume that any Europan biota would have to be essentially anaerobic and prokaryotic. In the lines that follow, some of the anaerobic microorganisms which could probably survive and grow in a Europan habitat are considered.

Photosynthetic organisms can take inorganic carbon in the form of CO_2 and fix or assimilate it into a usable organic form such as carbohydrates. Organisms with this ability are considered autotrophs or "self-feeders." Conversely, organisms ranging from humans to microorganisms that cannot fix inorganic carbon and thus are forced to consume organic carbon in the form of preformed biochemical compounds are called heterotrophs. Today's heterotrophs accomplish this by consuming plants and animals; however, at the onset of biotic history it is believed that primitive organisms survived by consuming the abiotically formed organic molecules present in the "Oparin broth." These organisms, believed to be the first organisms on Earth, had neither autotrophic nor photosynthetic capabilities. Only when these early cells began to exhaust the oceanic supply of nutrients did a need arise for autotrophic capability and eventually the ability to use light as an energy source to fix carbon. Therefore, the first anaerobic microorganisms were probably ancestors of the early heterotrophic prokaryotes that later developed autotrophic capabilities. Some of the fermentative *Clostridium* species have been suggested as such early heterotrophs. These organisms derive their energy by the anaerobic fermentation of many different organic compounds. Some of the sulfate-reducing bacteria which are strict anaerobes are also known to be capable of fermentative growth in the absence of sulfate. However, since *Clostridium* and sulfate reducers are highly evolved eubacteria, they cannot be considered as the earliest heterotrophic ancestors.

The study of phylogenetic trees determined from 16S rRNA sequence comparisons has led to the establishment of three major taxonomic groupings or fundamental cellular kingdoms, namely, eukaryotes, eubacteria, and archaebacteria. The universal ancestor was probably different from any of these three groups of organisms. However, some of the archaebacteria, particularly, the extreme thermophiles (e.g., *Thermoproteus, Pyrodictium, Sulfolobus, Desulfurococcus*), sit relatively close to the intersection of the three primary phylogenetic lines. Therefore these extreme thermophilic archaebacteria may more closely resemble some of the most ancient prokaryotes. In fact, some of them are fermentative anaerobes that have the ability to use sulfur, instead of oxygen, in their respiratory mode of metabolism.

The extreme thermophiles live at high temperatures (up to 110°C) and are very versatile microorganisms. For instance, the *Thermoproteus* species can grow in a strictly chemolithoautotrophic manner by deriving energy from the oxidation of inorganic compounds such as H_2 by means of sulfur instead of oxygen and utilizing CO_2 as its sole carbon source. Alternatively, they can grow by sulfur respiration of different organic substrates. The products of their fermentation during heterotrophic growth are CO_2 and H_2S. Instead of sulfur, malate can serve as a terminal electron acceptor. Thus the same organisms can be fermenters (heterotrophs) or chemolithotrophs (autotrophs), depending on the environmental conditions. These organisms represent possible bridges between fermentative metabolism and sulfur-dependent autotrophism or chemolithoautotrophism.

The importance of sulfur in ancient organisms has also been emphasized. Using several different methods of analysis of rRNA sequences, it has been found that sulfur metabolism is primitive for 98% of the possible phylo-genetic trees. These sulfur metabolizers (e.g., *Thermoproteus tenax*) have been classified as some of the most ancient microorganisms. Indeed, specimens of three extreme thermophilic archaebacteria (with morphologies similar to *Thermoproteus, Thermofilum,* and *Pyrodictium*) have been isolated using the *Alvin* submersible from the inside of a hydrothermal vent (at 130°C nominal temperatures) located at the Juan de Fuca Ridge in the Pacific Ocean about 500 km west of Seattle. The *Pyrodictium* gen. nov., is a new genus of submarine sulfur-reducing archaebacteria which grows optimally at 105°C.

In conclusion it appears that the earliest heterotrophic and autotrophic ancestors could have inhabited geothermal sulfur springs on the primitive Earth. Furthermore, one could not rule out the possibility of the existence around Europan deep oceanic vents of bacterial ecosystems similar to those existing around the Earth's deep-sea hot springs. Table 6-5 indicates the basic reactions characteristic of these heterotrophic and autotrophic microorganisms.

Table 6-5: Anaerobic Habitability of Europa by Archaebacteria—Energy-Yielding Metabolic Reactions

Metabolism	Reactions yielding energy and biosynthetic products (B.P.)	Example(s)
A. Thermophilic archaebacteria		
Heterotrophic:		
Fermentation	Yeast extract $\rightarrow CO_2$ + B.P.	*Staphylothermus marinus*
Sulfur respiration	Organics* + S $\rightarrow H_2S + CO_2$ + B.P. *(sugars, alcohols , formate, acetate)	*Thermoproteus tenax* (1) *Desulfurococcus mobilis*
Autotrophic:		
Sulfur reduction (S/H autotrophy)	$CO_2 + H_2 + S \rightarrow H_2S$ + B.P. (CO_2 assimilation via reductive TCA cycle)	*Pyrodictium occultum* *Thermoproteus neutrophilus Sulfolobus ambivalens* (2)
B. Methanogenic archaebacteria (3)		
Heterotrophic:		
Fermentation	Organics* + $H_2 \rightarrow CH_4$ + B.P. *(CH_3OH, CH_3NH_2, formate, acetate)	Many methanomicrobiales
Autotrophic:		
CO_2 reduction by H_2 (4)	$CO_2 + H_2 \rightarrow CH_4$ + B.P.	All methanobacteriales and methanococcales

(1) Facultative autotrophs.

(2) May also grow aerobically oxidizing S° to SO_4.

(3) They also reduce S° to H_2S.

(4) Some use Fe° instead of H_2 as the only source of electrons.

Table 6-6: Habitability of Europa by Photosynthetic Eubacteria—Energy-yielding Metabolic Reactions

Metabolism	Reactions yielding energy and biosynthetic products (B.P.)	Example(s)
A. Green non-sulfur bacteria		
Typical anaerobic photoheterotrophs (1)	Organics $\xrightarrow{light}$ Biosynthetic Products	*Chloroflexus aurantiacus*
	Also facultative photoautotroph (fixes CO_2) and aerobic chemoheterotroph	
B. Green sulfur bacteria (anoxygenic)		
Stricly anaerobic photoautotrophs (1, 2)	$CO_2 + 2H_2S \xrightarrow[P(I)]{light} (CH_2O) + S_2 + H_2O$	*Clorobium limicola*
	CO_2 assimilation via reductive TCA cycle.	
C. Cyanobacteria (oxygenic and facultative anoxygenic)		
Typical oxygenic photoautotrophs (3,4)	$CO_2 + H_2O \xrightarrow[P(I \& II)]{light} (CH_2O) + O_2$	*Oscillatoria*
	All have Calvin-Benson cycle	
Facultative anoxygenic H_2S photoautotrophs (4)	$CO_2 + 2H_2S \xrightarrow[P(I)]{light} (CH_2O) + S_2 + H_2O$	*Oscillatoria limnetica* (5)
	Similar to green sulfur bacteria	

(1) Similar to plant photosystem I: uses bacteriochlorophylls (Bchls) ferrodoxins (iron-sulfur proteins) and NADP carriers.

(2) They can photoassimilate acetate but only if H_2S and CO_2 are simultaneously present. CO_2 and acetate are assimilated by reversal of the tricarboxylic cycle (reductive TCA).

(3) They may survive aerobically in the dark at the expense of the dissimilation of glycogen to glucose and the oxidative pentose phosphate cycle. They have no functional TCA cycle.

(4) Similar to plant photosystems I and II: use chlorophyll a.

(5) *Oscillatoria limnetica* is a halophilic cyanobacteria capable of anaerobic, sulfide-dependent photoassimilation of CO_2. Sulfide inhibits photosystem II and induces an enzyme that allows sulfide to donate electrons to photosystem I. Elemental sulfur accumulates extracellularly. In the dark ATP can be generated from stored polyglucose reserves by either of two means: anaerobic respiration using the accumulated elemental sulfur as the electron acceptor; or a homolactic fermentation. This metabolism is similar to that of green sulfur bacteria and is widely distributed among non-heterocystous filamentous cyanobacteria.

It is believed that as early heterotrophic life on Earth established itself, the accompanying continuous decrease of fermentable substrates acted as a further evolutionary pressure toward autotrophy. Another closely related group of autotrophic anaerobes that do not require the use of light are the methanogenic archaebacteria which may also resemble some of the most ancient microorganisms on Earth. They obtain their energy and biochemical substrates simultaneously from the reduction of CO_2, by hydrogen, or other reduced compounds. The methanogens are also able to reduce elemental sulfur to H_2S by means of H_2 and obtain energy from this process similar to the extreme thermophiles. Representatives of the above two groups of archaebacteria, the anaerobic sulfur chemolithotrophs and the methanogens, have been isolated from the bottom layers of the Black Sea. Therefore organisms such as these could in theory be able to survive in the bottom layers of a Europan ocean. From a terrestrial evolutionary point of view, the picture that emerges by comparing the energy metabolisms of the sulfur thermophiles and the methanogens as well as their 16S rRNA sequences and their relative positions in the phylogenetic tree is that the ancestral archaebacterium was in essence an extreme sulfur-dependent thermophile which eventually gave rise to the methanogenic archaebacteria. Although the evolutionary connection of the thermophilic and methanogenic archaebacteria with the anoxygenic green sulfur phototrophic eubacteria is certainly more distant, there are also some similarities in their energy metabolisms which indicate an evolutionary relationship. Compare tables 6-5 and 6-6.

On the Earth the role of bacteria involved in the sulfur cycle was also very important during the subsequent phases of evolution. In fact, they are supposed to have been the originators of anoxygenic photosynthesis carried out by green sulfur bacteria in anaerobic environments. The earliest phototrophic bacteria used H_2S as a reductant (table 6-6.). They performed only one light conversion reaction, mediated by bacteriochlorophyll (photosystem I), and had a simple electron transport chain that included ferredoxins and cytochromes. They fixed CO_2 via a reductive tricarboxylic acid cycle. The electron donor for photosynthesis was H_2S, which was oxidized to extracellular elemental sulfur. These green photosynthetic bacteria, represented by extant Chlorobiaceae, are the strict anaerobes presumed to be the ancestors of blue-green bacteria or cyanobacteria. On the basis of 16S rRNA sequence analysis, a possibly more primitive group of phototrophic eubacteria is the green non-sulfur bacteria (anaerobic photoheterotrophs) which utilize organic compounds in a fermentation mode by using light energy for the generation of proton gradients and electron transport (table 6-6.).

Habitability by Aerobic Life

The photosynthetic cyanobacteria are characterized by having two photosytems (I and II) and using H_2O as a reductant instead of H_2S. The photosynthetic process of cyanobacteria is very similar to that of higher plants. They produce oxygen and are aerobic prokaryotes. Considering the limited amount of light available on Europa, the development of the anoxygenic photosynthesizers similar to those mentioned in the previous section would probably be somewhat limited if not altogether unlikely. It is therefore very difficult to answer the question of whether the envisaged putative Europan biological evolution would have subsequently led to the emergence of oxygenic photosynthetic microorganisms.

On the Earth, even though the presence of substantial amounts of oxygen in the atmosphere did not prevail until 2 billion years ago, the cyanobacteria may have appeared much earlier as indicated by some of the affinities of the microfossils found in Africa and Australia, the latter being 3.5 billion years old. It is possible however that the organisms represented by these microfossils had not yet developed the capability to utilize H_2O as a reductant and were still utilizing in its place the H_2S that their presumed ancestors, the green sulfur photosynthetic bacteria, had been using earlier. Indeed, some species of cyanobacteria, such as *Phormidium frigidum* which is found in ice-covered Antarctic lakes and *Oscillatoria limnetica* which is found in hypersaline lakes, are adaptable to extreme environments and can live under anoxic conditions. Furthermore, the latter eubacterium, *O. limnetica*, can perform anoxygenic photosynthesis at the expense of hydrogen sulfide as a hydrogen donor.

Even though it is highly speculative, it is possible that such an evolutionary transition from anoxygenic photosynthesizers to oxygenic photosynthesizers could have taken place on Europa. If this step did in fact occur, the upper part of the Europan ocean would have subsequently contained oxygen gas from the molecular oxygen released by the cyanobacteria. The partial saturation of the ocean waters could in turn bring important changes in the oceanic life, such as the possibility of establishing ecologically cooperative associations with anaerobic organisms living at lower levels of the Europan ocean as it happens in ice-covered Antarctic lakes. The above discussion on the habitability of Europa by photosynthetic microorganisms is summarized in table 6-6.

The low-temperature habitats of Earth's oceans and lakes maintain a host of aerobic and anaerobic ecosystems which provide an ideal starting point when one attempts to create a model for a speculative Europan biota. The ecosystems formed by the microbes that live under the ice in the cold waters of the Arctic and Antarctic oceans are of particular interest. The microorganisms inhabiting the perennially ice-covered Antarctic lakes have been extensively studied. The structural features and dynamics of these lakes have been described as a general model for Martian paleolakes. By changing the scale dramatically they could also be used as models for the subsurface ocean of ice-covered Europa.

The ecosystem of the Antarctic lakes is characterized by ecologically cooperative associations between several species of pennate diatoms, heterotrophic bacteria which occur abundantly throughout much of the benthic regions of these lakes, and microbial mats composed primarily of blue-green bacteria, specifically the cyanobacteria *Phormidium frigidum* and *Lyngbya martensiana*. These microbial mats are precipitating calcite and trapping and binding sediment, forming alternating laminae of organic and inorganic matter. A common feature of many of these benthic mats is their development into modern, cold water stromatolites. Because of the unique nature of these lakes, where the habitats vary chemically from fresh water to saline and from oxic to anoxic conditions, these stromatolites have been suggested as analogs of those that once inhabited deep water Precambrian systems. One of the more interesting features of the microorganisms living in the Antarctic lakes is not only their adaptation to the cold temperatures but their ability to live adaptatively in either aerobic or anaerobic conditions. Therefore, a speculative suggestion is that these organisms might possibly be able to survive and grow, to a limited extent, in a Europan ocean.

Conclusions

The observed abundance of H_2O in the satellites of the outer solar system could support the development of life as we know it if the liquid state can be maintained. We believe this could be the case of Europa as shown by the data from the Voyager flybys of Jupiter and the studies made on this satellite by two of us as well as other investigators.

We have presented a physical and chemical model for the possible habitability of Europa based upon studies of the evolution of the Galilean satellites and a comparison of these satellites with the chemical composition of carbonaceous chondrites. We conclude that Europa has amounts of reactive organic or inorganic compounds of the order of 10^{21} g or better, as shown in table 6-4. These compounds, being made of the different biogenic elements (H, C, N, O, S, P), are sufficient to provide the essential biochemical molecules for the emergence of life. In addition to water and biochemical molecules there are energy sources available on Europa, such as sunlight near the surface, and, more importantly, submarine hot springs at the floor of the Europan ocean, which, like the volcanoes on Io's surface and the terrestrial suboceanic vents, contain substantial amounts of sulfur and sulfur compounds.

Since energy and the necessary biochemicals may have been present, life could have emerged on Europa if the early conditions of this satellite would have been favorable. Whether such a process took place or not, the Europan environment could have provided, and may provide today, the necessary conditions for the survival and growth of many of the anaerobic microorganisms which thrive in terrestrial deep oceanic vents. Among the prokaryotic microorganisms capable of surviving and growing on the bottom sediment layers of Europa's ocean are the sulfur-dependent thermophilic archaebacteria and their evolutionary descendants, the methanogenic archaebacteria.

We also conclude, as other investigators have, that tidally heated habitable zones could possibly exist, with even more favorable conditions than Europa, around giant planets in planetary systems beyond the solar system.

These studies emphasize the importance of the Galileo Jupiter Orbiter Mission for closer observations of Europa. Observations that could be performed by Galileo include monitoring of Europa to look for the vapor and frost cloud that could result from a fracturing event, and high resolution imaging of the surface to elucidate the processes involved in forming the linear features as well as the variations in their albedo and coloration. Furthermore, radar sounding could be used to detect if liquid water is present. It is also suggested that serious consideration should be given to future missions using landing spacecraft that could penetrate the surface and perform *in situ* surface and subsurface measurements on Europa.

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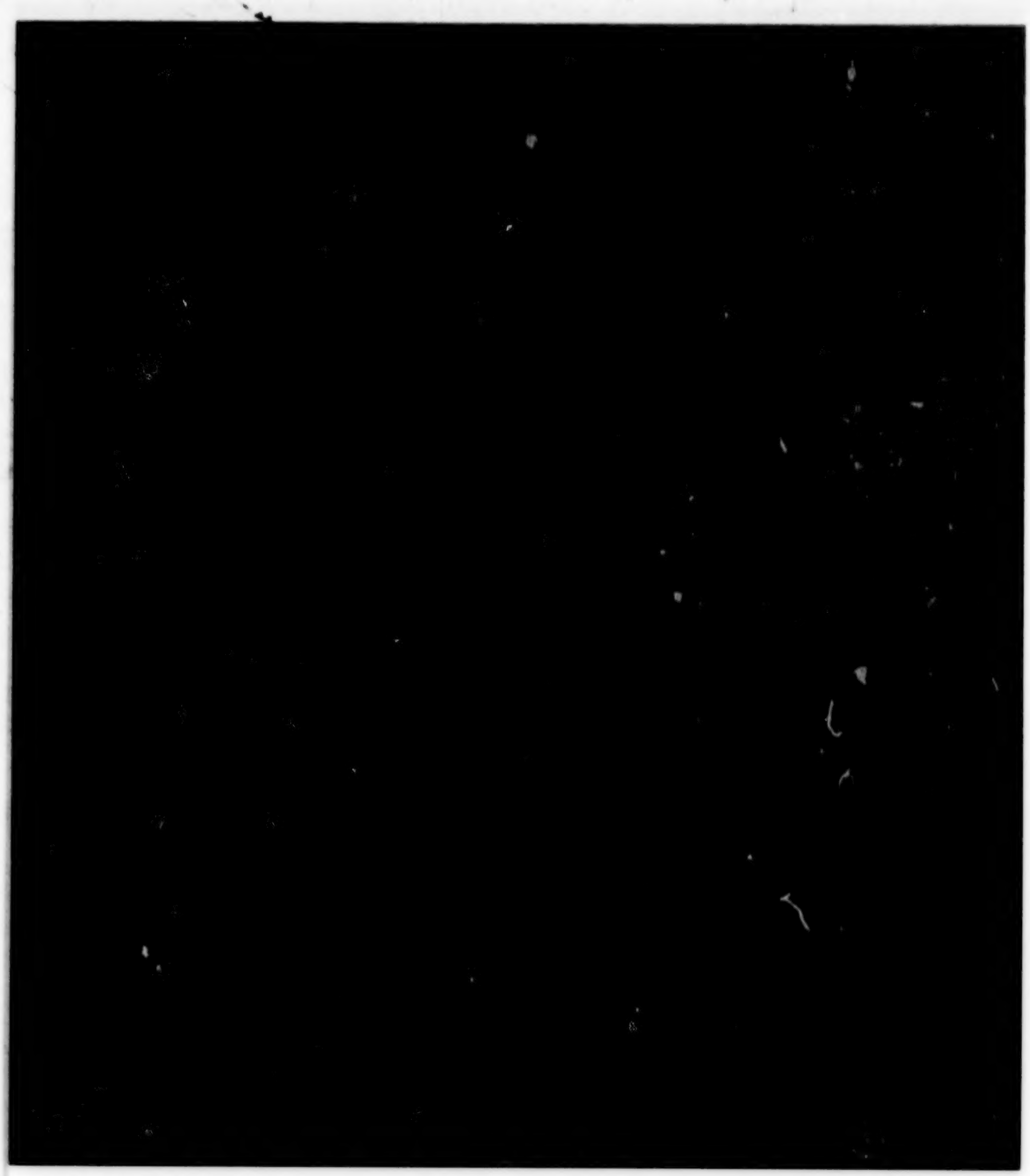

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Titan

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T. Owen, D. Gautier, F. Raulin, and T. Scattergood

Titan is Saturn's largest satellite, with a diameter greater than that of Mercury and an atmosphere denser than Earth's. This atmosphere began as an assemblage of gases produced by the outgassing of the materials that accreted to form the satellite. Chemical reactions during the accretion process and over the subsequent eons produced the gases that we find today. In fact, the process is far from over: Titan offers us an enormous natural laboratory in which a variety of experiments in chemical evolution are taking place today.

We are, therefore, confronted with a kind of paradox. Here is a solid body, large enough to retain an atmosphere and small enough to allow hydrogen to escape, leading us to expect to find the same oxidizing conditions we encounter on the inner planets. Yet, unlike the atmospheres of those familiar bodies, Titan's atmosphere is highly reducing, consisting primarily of nitrogen, with methane the next most abundant reactive constituent.

ORIGINAL PAGE
COLOR PHOTOGRAPH

The surface pressure is 1.5 bars and the surface temperature is close to 94 K. It is this low temperature that explains the paradox, since under these conditions water is totally frozen out so there is no readily available source of oxygen to change the character of the atmospheric chemistry. Even though methane is continually being dissociated and the resulting hydrogen escapes from Titan's exosphere, the reducing state of the atmosphere is maintained.

Nevertheless some CO and CO_2 are present, in addition to a large number of hydrocarbons, nitriles, and complex polymers that form a ubiquitous smog in Titan's lower stratosphere. Methane should condense to form hazes of crystals in the upper troposphere, while ethane clouds may exist at lower altitudes. Indeed, it is possible that sufficient ethane has been produced to form lakes, seas, or even a global ocean on the satellite's surface.

The chemistry that produces these various compounds has not yet been elucidated in detail. Is the CO primordial, or is it produced by reactions with infalling ice? What is the relative significance of magnetospheric electrons, galactic cosmic rays, and solar ultraviolet light in driving the chemical reactions? What level of chemical complexity has been achieved? For example, are there polymers of HCN? Is adenine being produced? What relationship does this chemistry have to the various pathways postulated for prebiotic chemistry on the primitive Earth?

In this chapter, we shall review the current status of the analysis of Voyager and ground-based observations of Titan and the interpretation of these data in terms of laboratory experiments and theoretical models. We will conclude with a description of a proposed mission to advance our knowledge and some considerations of the significance of Titan from a planetary system perspective.

Observations of the Atmosphere

Our knowledge of the structure and the composition of the atmosphere of Titan comes primarily from observations made by Voyager 1 in November 1980. The spacecraft passed Titan at a distance of 4000 km from its surface, making a wide variety of measurements with all of the instruments it carried.

Table 7-1 gives a summary of the main physical characteristics of the satellite. The voyager radio occultation experiment achieved the first

Table 7-1: Characteristics of Titan

Surface radius	2575 km
Mass	0.022 Earth mass
Surface gravity	135 cm s^{-2} (0.14 Earth gravity)
Mean density	1.881 g cm^{-3} (0.34 Earth density)
Distance from Saturn	1.226×10^6 km = 20 R_S
Orbit period around Saturn	15.95 day
Orbit period around Sun	29.46 years
Obliquity	26°.7 (assumed equal to Saturn's)
Temperature	
Surface	94 K
Tropopause (42 km)	71.4 K
At 200 km altitude	~170 K (at the equator)
Surface pressure	1496 mbar

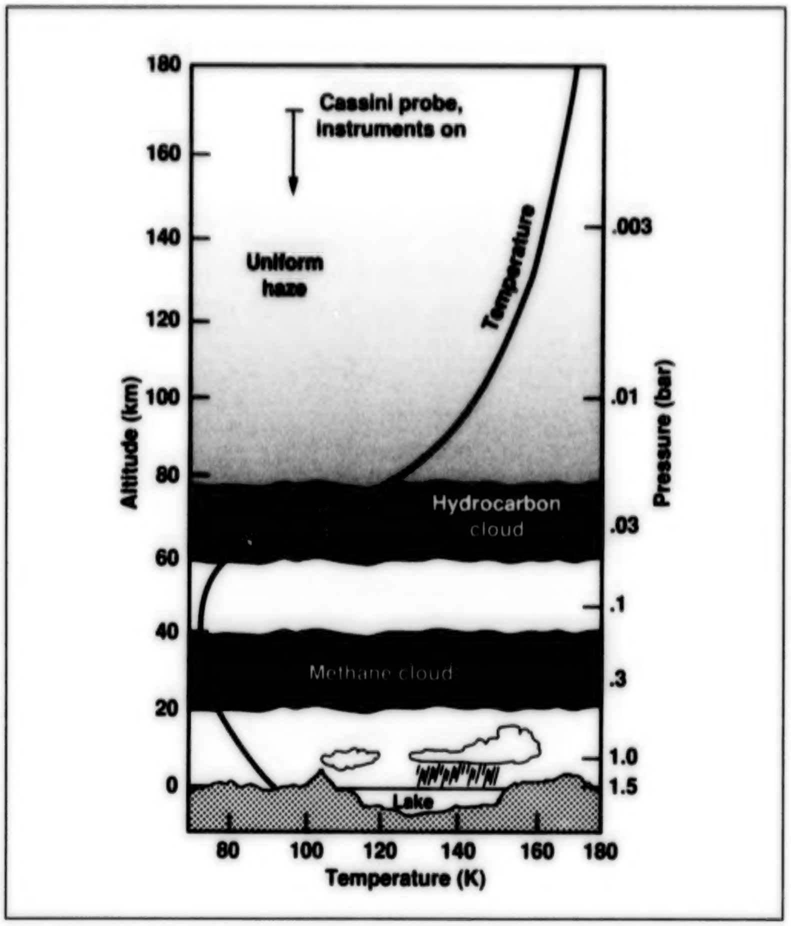


Figure 7-1. The temperature structure of the Titan atmosphere from early Voyager results. This approximate thermal profile has just been updated (see text); the landscape is drawn to demonstrate some of the possibilities that our present ignorance permits.

measurement of the radius of the solid surface of Titan which is hidden from view by layers of smog in the atmosphere. The optical limb of Titan, as defined by the smog, is at an altitude of 200 km, but layers of aerosols have been detected up to 500 km. The radio occultation experiment, combined with results from the Voyager infrared and ultraviolet spectrometers, indicates that the main atmospheric component is molecular nitrogen. These data also provide a description of the vertical thermal structure of the atmosphere at Titan's equator. A recent reanalysis of the radio occultation data indicates that the surface temperature of Titan could be between 92.5 and 101 K, the tropopause temperature between 71.4 and 74.5 K, and the stratospheric temperature between 164 and 172 K around the 160 km altitude level. The temperature profile above the 200 km altitude level is not known but is probably isothermal or weakly increasing with altitude at least up to 300 km (fig. 7-1).

A major part of the uncertainty identified in the reanalysis originates from lack of a precise value for the atmospheric mean molecular weight. As much as 20% of the atmosphere might consist of non-radiogenic argon, a mixture of isotopes with atomic masses 36 and 38. Argon is expected to be present for cosmogonical reasons, as we shall see below. Unfortunately, this gas is very difficult to detect directly from a remote vantage point. The only indication of its existence at present is a possible contribution to the mean molecular weight of the atmosphere. The present uncertainty in this quantity precludes any firm conclusions (see table 7-2).

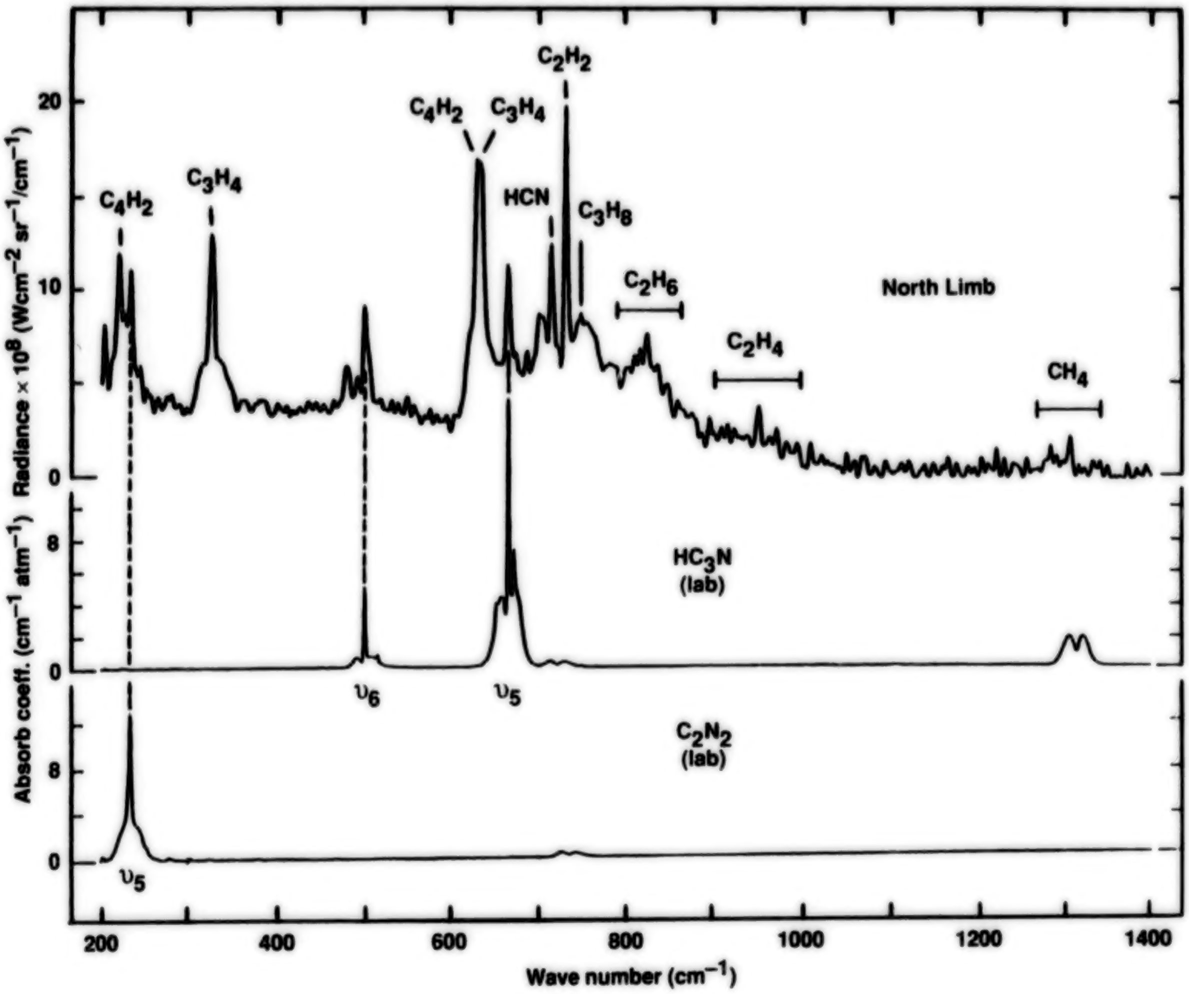
Table 7-2: Stratospheric Composition of Titan

Gas	Mole fraction	Comments
Major components		
Nitrogen (N_2)	0.757–0.99	
Argon (Ar)	0–0.21	Inferred indirectly
Methane (CH_4)	0.005–0.034	May condense in the troposphere (0.002–0.21 at the surface)
Hydrogen (H_2)	0.002–0.006	
Hydrocarbons		
Ethane (C_2H_6)	1×10^{-5}	At the equator
Acetylene (C_2H_2)	1.8×10^{-6}	At the equator
Propane (C_3H_8)	7.0×10^{-7}	At the equator
Ethylene (C_2H_4)	5.0×10^{-9}	At the equator
Diacetylene (C_4H_2)	10^{-8}–10^{-7}	At the north pole
	1.4×10^{-9}	At the equator
Nitriles		
Hydrogen cyanide (HCN)	1.4×10^{-7}	At the equator
Cyanoacetylene (HC_3N)	10^{-8}–10^{-7}	At the north pole
Cyanogen (C_2N_2)	10^{-8}–10^{-7}	At the north pole
Oxygen compounds		
Carbon dioxide (CO_2)	1.1×10^{-8}	Assuming a constant mixing ratio
	3.3×10^{-9}	At 108 km altitude level, assuming a distribution parallel to the one of Samuelson et al. (1983)
Carbon monoxide (CO)	6×10^{-5}	In the troposphere
	$<4 \times 10^{-6}$	In the stratosphere

Methane is clearly present in the stratosphere as indicated by its features in the infrared spectrum. In fact, this gas was detected spectroscopically in the troposphere 45 years ago with the 2.1-m reflector of the McDonald Observatory. The CH_4 stratospheric molar abundance is between 0.5 and 3.4%. Near-infrared ground-based measurements unfortunately do not provide accurate estimates of the CH_4 abundance in the troposphere. It is generally considered, although not on a firm basis, that the CH_4 abundance at the surface is around 5-10%. Since CH_4 may condense at the tropopause, its tropospheric mixing ratio cannot be deduced from the stratospheric value, except if the latter is less than the tropopause saturation value. The lower limit in the troposphere is then 0.5%. If argon is included in maximal proportions, the upper limit of the CH_4 molar abundance at the surface may be as high as 21%; if argon is absent, the maximum value is reduced to 11.7%. The stability of the atmosphere requires unsaturated relative humidities in the lower atmosphere; CH_4 clouds are expected somewhere between the ground and the tropopause level.

The infrared spectra recorded by the Voyager IRIS (Infrared Imaging Spectrometer) experiment have also revealed the presence of hydrocarbons and nitriles in the stratosphere of Titan (fig. 7-2 and table 7-2). These species are formed from the dissociation of CH_4 and N_2. The stratospheric abundances of C_2H_2, C_2H_4, C_2H_6, C_3H_4, C_3H_8, C_4H_2, and HCN at the equator have recently been revised taking advantage of new improvements to the determinations of the equatorial thermal profile and of more precise laboratory measurements of band intensities and other spectroscopic parameters. Most of the planetary emission in the observed bands originates from about the 108 km altitude level, except for the ν_5 band of C_2H_2 which is formed at an altitude around 130 km. (Note that all stratospheric minor components, except CO, condense below 50 to 70 km altitude levels.) The abundances observed at these levels are systematically

Figure 7-2. A section of Titan's infrared spectrum as recorded by the Voyager IRIS instrument.



higher than predicted by detailed photochemical models except for C_2H_4 and CO_2 for which the agreement is satisfactory. This does not necessarily imply that present photochemical models must be revised, however. In the procedure used in these models to calculate the mixing ratios of various stratospheric species above the saturation level, the magnitude of the eddy diffusion coefficients in the lower thermosphere and in the lower stratosphere play a key role. A readjustment of these coefficients may permit the predictions from the models to be reconciled with the observations.

More constraints come from near horizontal viewing observations made by Voyager 1 at the north pole of the satellite. Spectra were recorded at various altitudes, allowing the retrieval of vertical distributions of stratospheric absorbers. Vertical distributions of C_2H_2, C_2H_6, C_3H_4, C_4H_2, HCN, and HC_3N have thus been obtained. All species exhibit mixing ratios increasing with height, at least up to the 300-400 km altitude level, except C_2H_6 for which the concentration holds fairly constant above the saturation level. Some distributions have shapes, but not amplitudes, compatible with photochemical models. That is the case for C_2H_6, C_4H_2, HC_3N, and C_3H_4, but not for C_2H_2 and HCN. It must be noted that the models refer to the whole disk of Titan while the north pole photochemistry may be peculiar because of the geometrical configuration in this region. That this peculiarity is not caused primarily by seasonal effects may be seen by a more detailed examination of the spatial distribution of stratospheric constituents.

No longitudinal variation in temperature or abundances was detected at the equator. On the other hand, HCN, HC_3N, C_2N_2, C_4H_2, and C_3H_4 appear to be strongly enhanced at the north pole compared to the equator. Moreover, comparison of IRIS spectra at the north and south poles suggests that abundances of nitriles are similar in the two regions, at least on the night side. That confirms the special nature of the photochemistry at the poles, but rules out explanations based on seasonal effects.

The case of CO is especially intriguing. It is unusual to find an oxygen-containing compound in this reducing environment, and the vertical distribution of CO in Titan's atmosphere is itself something of a puzzle. CO was discovered in the near infrared by ground-based observations. A CO/N_2 mixing ratio of about 6×10^{-5} was derived. The observed lines are believed to be formed mainly in the troposphere of the satellite. Since CO does not condense at Titan's temperatures, its mixing ratio is expected to be constant with height up to very high atmospheric levels. However, recent millimeter heterodyne observations lead to the conclusion that the stratospheric mixing ratio cannot exceed 4×10^{-6}. The only way to reconcile the near-infrared and microwave observations is the possibility that CO is severely depleted in the stratosphere compared to its abundance in the troposphere. Such behavior is in conflict with presently available models of Titan that consider CO to be uniformly mixed throughout the whole atmosphere. The main difficulty is how to maintain a vertical gradient in the presence of even a minimal vertical mixing, considering the stability of the CO molecule.

The CO that we find on Titan today may either be original or formed by reactions with H_2O that enters the atmosphere from the outside in the form of ice crystals or meteoritic dust. A similar ambiguity exists for the production of CO_2. This molecule is produced from CO and OH, mainly through the reaction

$$CO + OH \rightarrow CO_2 + H \qquad (1)$$

Once again the oxygen (in the form of OH) can either come from ice:

$$H_2O + h\nu \rightarrow H + OH \qquad (2)$$

or it could be contributed from the (primordial) CO:

$$CO + e \rightarrow C + O(^1D) + e \qquad (3)$$

$$O(^1D) + CH_4 + CH_4 \rightarrow OH + CH_3 \qquad (4)$$

where $O(^1D)$ is the excited state of atomic oxygen.

Another perspective on these possibilities is furnished by the evidence for an enhanced value of D/H in Titan's atmosphere. The abundance ratio CH_3D/CH_4 on Titan seems to require the preservation of an interstellar value of D/H in the volatiles incorporated in the satellite, because processes acting since the formation of Titan will not produce the observed enrichment. If this is indeed true, then one might expect some residue of the interstellar dominance of CO over CH_4 to remain in the gases produced during the accretion of Titan. In fact, the methane we find on Titan today may not be interstellar methane at all, but rather the result of the accretional heating (during Titan's formation) of organic compounds trapped in interstellar grains. Whether or not the CH_4 is original, the high D/H suggests the CO may well be, in which case one would expect an amount roughly equivalent to the N_2 abundance we find today in the atmosphere. Reactions 1-4 will cause the irreversible conversion of CO to CO_2 which we should find deposited on the surface today.

Laboratory Simulations and Theoretical Models of Titan's Atmosphere

What we know about the composition of Titan's atmosphere is certainly incomplete. In the absence of detailed, *in situ* measurements of the actual constituents and energy source, investigators have turned to laboratory experiments and computational models that simulate the physics and chemistry that are believed to be taking place. Each of these approaches has its limitations but both can be used separately, or preferably, in concert to make predictions about the processes responsible for the observed atmospheric species (table 7-2) and about species not yet observed, but which should be looked for by future probes sent to Titan (table 7-3).

Table 7-3: Upper Limits on Abundance of Nitriles Expected to be Present in Titan's Atmosphere

Nitriles	Mole fraction
Acetonitrile (CH_3–CN)	$<2.5 \times 10^{-7}$
Proprionitrile (C_2H_5–CN)	$<5.0 \times 10^{-7}$
Acrylonitrile (CH_2=CH–CN)	$<2.0 \times 10^{-7}$
Crotononitrile (CH_3–CH=CH–CN)	$<4.0 \times 10^{-8}$
Allylcyanide (CH_2=CH–CH_2–CN)	$<4.0 \times 10^{-8}$
Methacrylonitrile (CH_2=C(CH_3)–CN)	$<7.5 \times 10^{-8}$
Cyanopropyne (CH_3–C=C–CN)	$<2.5 \times 10^{-7}$

An important example of the deductions that have been made indirectly (which will require confirmation) is the possibility of an ethane ocean. Such an ocean is postulated to cover much, if not most, of Titan's surface and to contain compounds formed in the atmosphere. The existence of this ocean is based upon both theoretical modeling and laboratory studies of the photochemistry of methane, the most abundant hydrocarbon in Titan's atmosphere. Such studies suggest that ethane will be the predominant product of methane photochemistry and will condense at the low temperature prevailing over Titan's surface. More details about this putative ocean will be given below.

Most of the laboratory studies that have been done thus far to examine the chemistry and physics of Titan's atmosphere can be divided into two broad classes. The first class consists of experiments that are carried out to provide specific data needed to interpret spectroscopic and optical observations of Titan or for inclusion in theoretical models of atmospheric physics and chemistry. Examples would be laboratory spectra of gases to identify features in Titan's spectrum and the determination of rate constants for specific chemical reactions.

The second class of laboratory experiments is more general in scope. These are attempts to simulate various types of chemical processes expected to occur in different regions of Titan's atmosphere. Most of these experiments have been done for the purpose of developing reasonable scenarios for explaining the compounds that have already been detected on Titan. In these studies, mixtures of CH_4, N_2, NH_3, and H_2 which represent models of Titan's atmosphere are subjected to various sources of energy, including ultraviolet light, electric discharges (simulating lightning and plasmas associated with meteor infall), and energetic electrons or protons representing those trapped in radiation belts around Saturn. After exposure of the gas mixtures to the appropriate energy source, the products are separated from the initial reactants and their chemical identities and yields are determined. These yields can then be compared to the amounts of the various species found on Titan in order to assess the roles that different chemical processes play in the chemistry of the atmosphere.

For example, when mixtures of N_2 and CH_4, in proportions similar to those found on Titan (table 7-2), were subjected to bombardment by energetic electrons (simulating processes in the upper atmosphere) or to electric discharges (hypothesized to occur in the troposphere), all of the nitriles and hydrocarbons detected on Titan were formed. Oxygen compounds and argon have not yet been included in these experiments. In addition, many other compounds including C_3, C_4, and C_5 hydrocarbons and nitriles were made. Certain of these compounds, for example acetonitrile (CH_3CN) and allene (CH_2CCH_2), were produced in amounts relative to HCN and C_2H_2, respectively, that are similar to those of some of the species already observed in Titan's atmosphere. Most of the nitriles exhibit far-infrared absorption bands so that it is possible to estimate the upper limits of their abundances from Voyager IRIS spectra as shown in table 7-3. These upper limits, compared with abundances of detected nitriles (HCN, HC_3N, and C_2N_2) are compatible with values expected from laboratory simulations. A number of complex organic heteropolymers could also be present.

On the other hand, the very idea has been challenged that magnetospheric electrons are responsible for the breaking of the N_2 bond (required to initiate formation of nitriles). The penetration depth of such electrons might be so small that they can only be involved in chemical processes occurring at high altitudes where organic molecules are not protected enough against photodissociation. Instead, galactic cosmic rays (GCR) could play a crucial role in these syntheses. Modeling studies suggest that the absorption of high energy GCR in the 0.1-50 mbar region of the atmosphere can generate energetic electrons that are able to induce several reactions driving the photochemistry. By coupling of the nitrogen and hydrocarbon chemistry, these processes can lead to the formation of HCN and other N-organics. Presumably reactions involving CO would also be initiated, although these have not yet been studied. This possibility is especially attractive in view of the mysterious deficiency of CO in the stratosphere (see above).

Finally, almost all of the experiments discussed above included the formation of solid/liquid products, often called tholins. These materials are generally reddish-brown to reddish-orange and may represent some of the substances that constitute the hazes and clouds that exist on Titan. Analyses of these materials show them to consist of a very complex mixture of hydrocarbons, nitriles, and other compounds but a definitive confirmation of the composition of any laboratory tholins has yet to be achieved.

In addition to laboratory simulations, modeling of the physics and chemistry of planetary atmospheres can be accomplished by the use of theoretical computations. Mathematical expressions representing the detailed steps in a chemical scheme are combined to "simulate" the process under study. Examples of this approach applicable to Titan include numerous models of the photochemistry of the atmosphere (discussed below), models of the effects of high-energy particles on the chemistry of the upper atmosphere, models of the scattering of light by aerosols (helpful in understanding the nature of the light that is reflected back to us from the satellite), and models describing the ways in which energy is transmitted and radiated by the atmosphere. Only the first two types of models will be discussed here as these explicitly involve the chemistry that can explain the species observed on Titan and those that may be present but not yet found.

Theoretical studies on the photochemistry of C_1 and C_2 hydrocarbons were first done in the mid-1970s. A model was developed several years later for the photochemical conversion of NH_3 (which may have been the initial form of nitrogen on Titan) into the N_2 found today. We have already referred to the very comprehensive and detailed model that includes all of the important features of previous work developed in the mid-1980s. This model includes the photochemistry of all of the hydrocarbons, nitriles, and oxygen compounds listed in table 7-2, plus the chemistry initiated by the influx of energetic particles from the magnetosphere, but not the effects of GCR. Some 150 elementary reactions are included, yielding a model with both advantages and disadvantages over the simpler models. The inclusion of such a large number of reactions reduces the chances of inadvertently leaving out important reactions, but the uncertainties in the results may be considerably increased owing to the need for rate constants for all of the reactions. Many of these reaction constants are only poorly known or are not known at all and must be estimated. This becomes especially critical when the particular reaction that controls the overall rate of the chain is either overlooked or poorly estimated. Even with these limitations, the photochemical models are quite useful in understanding the chemistry of Titan's atmosphere, in predicting compounds to search for, and in establishing the locations in the atmosphere in which to carry out such searches.

Summarizing the results of this model, all of the presently detected species in Titan's atmosphere can reasonably be accounted for by photochemical and energetic-particle driven reactions. However, the authors recognized that many of the reactions are only poorly understood and some uncertainties exist about certain details of the results. For example, a model suggests that large amounts of allene should be formed but none has been detected thus far. Also, as we pointed out above, this model in its original form does not give close agreement with the most recent evaluation of atmospheric abundances and distributions of other known species. Finally, except for the polyacetylenes, no compounds with more than three carbon atoms are included in the model, thus little can be said about the formation and nature of the known aerosols.

Photochemical models do indicate that large amounts of ethane (and some other organic compounds) should have been produced over Titan's history, thus leading to the idea that the satellite's surface may be covered by a global ethane ocean. The properties and ramifications of such an ocean will be discussed in the following section.

Endpoints of Atmospheric Chemistry: Aerosols and Oceans

As mentioned above, most of the compounds present in the gas phase must condense in the lower stratosphere because of the cold temperature. In this region, and around the tropopause, the concentration of condensable trace constituents in the gas phase must be very low. For example, all the nitriles, including HCN, have mole fractions smaller than 10^{-15} at altitudes ranging from 60 km down to 10 km.

The condensation of the low molecular weight organics would be induced by the submicrometer particles formed in the higher regions of the atmosphere through photopolymerization of C_2H_2, C_2H_4, and HCN. (The latter may actually polymerize on aerosol surfaces, rather than in the gas phase; a point that needs additional clarification.) These particles, of mean radius around 0.01 to 0.1 micrometers, would be responsible for the haze layers observed in the 200-400 km region. After precipitation down to the lower atmospheric layers, they would act as nucleation centers, allowing the condensation of lower molecular mass compounds. The resulting aerosol clouds would consist of particles covered by an external layer of more volatile organics and having a mean radius increasing with decreasing altitude.

The physical modeling of the processes that control the formation and vertical distribution of aerosols in Titan's atmosphere, as performed before the Voyager Mission, did not use the right atmospheric conditions. A new model has recently been developed which is based on a microphysical approach and uses an up-to-date atmospheric profile. It considers coagulation, condensation, diffusion, and precipitation processes, starting from a submicrometer particle in the high stratosphere. It includes the eventual vaporization of a part of the aerosol in the lower troposphere. Preliminary results obtained from this model indicate that there may be a permanent rainfall down to the surface of aerosol droplets with average diameters up to a few hundred micrometers. This is in agreement with the results of a recent multiple scattering radiative transfer model. Near the surface of Titan, a typical aerosol particle would consist of a core of several tens of micrometers which is mainly composed of nitriles and covered by an envelope consisting of C_1 to C_3 hydrocarbons. In these droplets, most of the organics would be highly enriched relative to the atmosphere. For instance, the concentration of C_2 and C_3 hydrocarbons would be three orders of magnitude higher in the aerosol close to the surface than those in the atmosphere.

Very close to the surface, most of the methane in the aerosol would be vaporized, feeding the lower troposphere with this compound. The resulting near-surface increase of the atmospheric methane mole fraction should lead to the condensation of this compound on Titan's surface, in order to maintain a quasi-thermodynamical equilibrium between surface and atmosphere.

Atmospheric methane is rapidly photolyzed, either directly in the mesosphere or catalytically through acetylene photolysis in the stratosphere. Given the first discovery in 1944 of CH_4 in Titan's atmosphere and repeated detection since, it seems very unlikely that we have been observing a very peculiar phase of Titan's history, where methane would have suddenly been abundant in the atmosphere. On the contrary, it can be assumed that methane must have been present in the current large amount in Titan's atmosphere during most of the history of this planetary body. This requires the existence of a CH_4 reservoir on Titan's surface.

The presence of a pure CH_4 ocean could play this role. However, previous analyses of data from the Voyager radio occultation experiment seem to indicate the absence of such an ocean, at least of a global ocean. But this does not rule out the presence of a methane-ethane ocean. In fact, this hypothesis appears to be the best model of Titan's surface at the present time.

Effectively, the major net product of methane photolysis is ethane. Because of the low atmospheric temperatures, ethane, like most other organics, condenses near the tropopause. It remains in a condensed phase down to low altitudes where it should be a liquid. Consequently, Titan's surface could be covered by an ocean of liquid ethane which could contain enough methane to act as the sought-for reservoir. The presence of methane in liquid ethane markedly decreases the atmospheric methane abundance and provides atmospheric conditions which are consistent with the observational data.

If we assume that the ethane flux downward to the surface has been roughly constant for the past 4.5 billion years and equal to the present photolytical rate of ethane production (5.8×10^9 cm^{-2} s^{-1}), then the minimum depth of the resulting ocean would be about 700 m. This ocean would be composed primarily of residual methane plus the ethane produced photochemically. It could be a global ocean, with this average depth, or it could be broken up into isolated lakes, depending on the topography. This latter scenario is consistent with recent radar measurements of Titan which show variations in "brightness" with rotation of the satellite. However, some have argued that Titan must either have a global ocean or no ocean at all, i.e., no large discrete bodies of liquid on its surface, to avoid changes in its orbit from tidal dissipation. We shall speak of this possible accumulation of liquid on Titan's surface as an ocean, for simplicity. This ocean should contain a noticeable fraction of dissolved gases (mainly nitrogen and, if present in the atmosphere, argon). It should also include additional solutes, in particular, organics coming from the atmosphere, either in the liquid or solid phases.

If one assumes that the ocean is in thermodynamic equilibrium with the adjacent atmosphere, knowing the surface temperature and the near-surface atmospheric composition, it is possible to calculate the bulk composition of the ocean. Using this approach, it is estimated that methane atmospheric mole fractions of 0.016 and 0.03 should correspond to methane mole fractions in the ocean of 0.13 and 0.25, respectively. In the latter case the mole fraction of dissolved nitrogen is about 0.05. Calculations on phase equilibria in the N_2–CH_4–C_2H_6 system under Titan conditions indicate that equilibrium between such a ternary liquid system and Titan's atmosphere at 94 K implies an atmospheric CH_4 mole fraction ranging from 0.095 to 0.2. In this case, the ocean composition varies from CH_4 (0.86) to N_2 (0.03), respectively. These calculations do not take into account the possible presence of argon. In addition they also ignore the likely presence of other solutes in the ocean.

Now the aerosols will continuously and irreversibly feed the lower troposphere, and finally the ocean, with ethane, but also with other organics. Few of these are in the liquid state at the temperature of the surface. But propane (C_3H_8) will be a liquid and is miscible in very large proportions in the methane-ethane ocean. If we assume that the flux of propane, like that of ethane, has been roughly constant during all of Titan's history and is equal to 1.4×10^8 cm^{-2} s^{-1}, then the mole fraction of propane in the ocean, $f(C_3H_8)$, is proportional to that of ethane. Thus, the main composition of the ocean should be CH_4–C_2H_6–C_3H_8–N_2–CO–(Ar), with $f(C_3H_8) = 0.024 \times f(C_2H_6)$. Most of the other compounds transferred to the ocean by means of the aerosols should be solid at the temperature of the surface. Depending on their density relative to the ocean they will either sink or float. In addition, depending on their solubility and their atmospheric flux to the surface, they must dissolve partly or totally. Thermodynamic modeling of these processes has recently been carried out, first by considering the case of a few of the low molecular weight organics and a C_2H_6–CH_4 solvent, neglecting the presence of dissolved N_2 and other gases, then by considering a C_2H_6–CH_4–N_2 solvent and a larger variety of solutes. The results suggest that all the non-dissolved solutes should be denser than the ocean even if its main constituents include argon and propane with high mole fractions (fig. 7-3). Thus the surface of Titan's ocean should be free of any organic icebergs.

The ocean would be very rich in organic solutes, with concentrations in the range 1 to 10^{-6} mole l^{-1}, depending on their chemical nature. With the exception of alkanes and cyclanes, which are very soluble in such a solvent, most of the other solutes, taking into account their plausible atmospheric

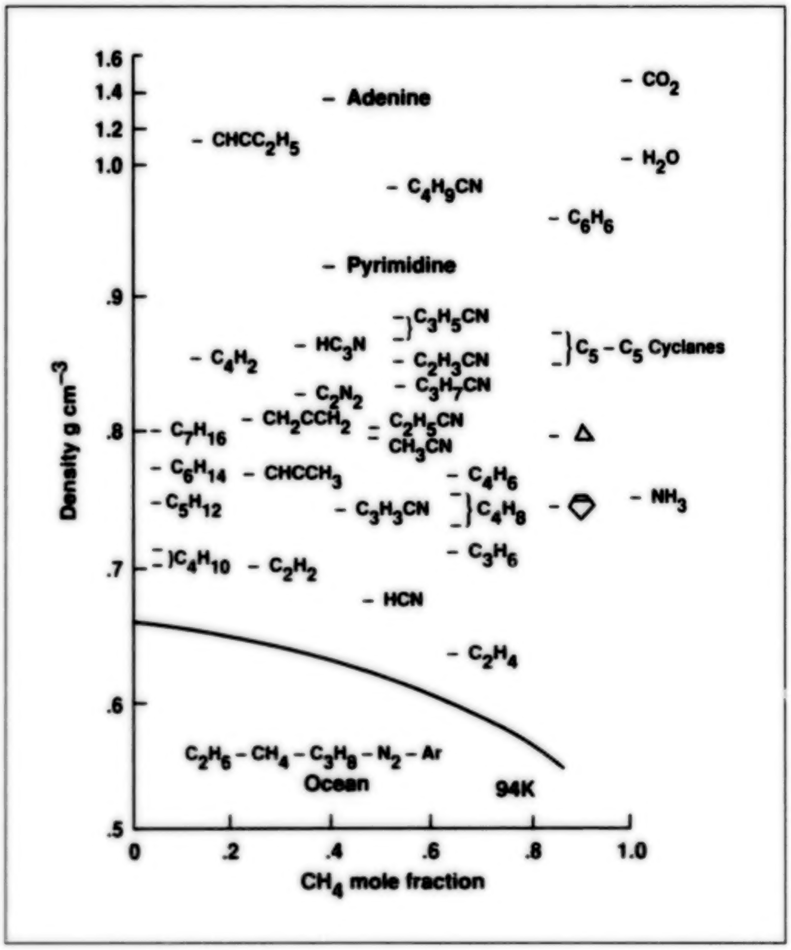


Figure 7-3. a) The curved line gives the density of the ocean as a function of its methane mole fraction (atmospheric argon mixing ratio = 0.17; oceanic propane mole fraction relative to ethane = 0.024). b) The labeled dashes show the densities of various possible solutes at the same temperature (94 K).

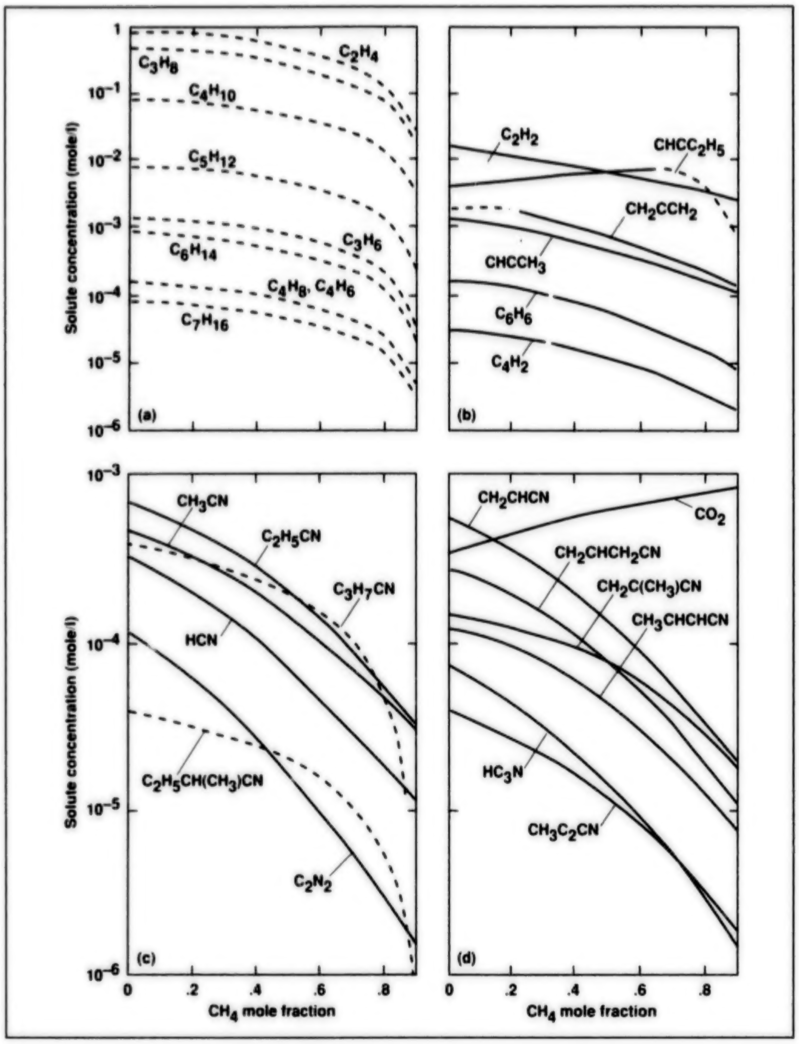


Figure 7-4. Estimated concentration of selected solutes as a function of methane mole fraction in the ocean (94K, argon free). a & b: hydrocarbons c: alkylnitriles; d: other nitriles and carbon dioxide. (Dashed line indicates that saturation is not reached).

flux, will reach saturation. This is the case, in particular, of acetylene, the solubility of which ranges between about 10^{-2} mole l^{-1}, corresponding to a mole fraction of about 4×10^{-4} (for an ethane-rich ocean), and 3×10^{-3} mole l^{-1}, corresponding to a mole fraction of about 10^{-4} (for a methane-rich ocean). For the nitriles, which are more polar than the alkynes and consequently less soluble in this apolar solvent, the solubilities are smaller. They range between 10^{-3} and 10^{-6} mole l^{-1}. For instance, the saturation concentration of HCN would be about 3×10^{-4} mole l^{-1} in an ethane-rich ocean, and 10^{-5} mole l^{-1} in a methane-rich ocean (fig. 7-4). The solubilities should be also very temperature dependent.

For the high molecular weight compounds the solubility must be very low, especially if the ocean is ethane rich. The only polymer slightly soluble in such a solvent seems to be polyethylene, and only in a methane-rich ocean. Thus a deep layer of organic compounds, mainly acetylene, HCN, and their oligomers or hetero-oligomers, should be present at the bottom of Titan's ocean. Such a reservoir of organics should buffer the oceanic concentration of these compounds.

But this milieu is not static. It should be a dynamic organic chemical environment. Its surface is continuously bombarded by galactic cosmic rays of very high energy (higher than a few 10 GeV) not absorbed by the atmosphere. Such particles could induce new reactions in this exotic ocean by processes analogous to those recently studied in the case of Triton. From this stage chemical evolution on Titan must follow a path very different from the terrestrial one: a chemical evolution in which liquid water is replaced by a cryogenic apolar solvent.

Titan and Exobiology

Despite this important difference from Earth, Titan remains an extremely interesting solar system object for exobiological studies. The interest obviously arises from the opportunity this satellite provides for the study of on-going organic synthesis under totally natural conditions over an immense span of time. The absence of liquid water also removes a readily available source of the oxygen that is required for many fundamental biochemical compounds, as we discussed above. Nevertheless the presence of CO, possibly as a primordially abundant gas, leaves open the possibility of more prevalent syntheses involving oxygen compounds in the past. Here the low temperature and reducing conditions are an advantage; unless there really is a global ocean, ancient materials should be well-preserved and readily accessible on Titan's surface.

Looking back over Titan's history, we note the possibility that ammonia was present during the early epochs and might even have been the precursor of the N_2 that we now find in the atmosphere. Hence, all of the simple molecules normally postulated as the necessary ingredients for prebiotic synthesis (CH_4, NH_3, CO, N_2, H_2O, and H_2) were probably available. There is even a possibility of more moderate temperatures (say 200 K) if sufficient ammonia were present to cause a large greenhouse effect. This would have been a short-lived condition since ammonia is so easily dissociated, but that process itself would have produced some interesting compounds as fragments combined with other radicals in the atmosphere and on the surface.

At the present time, however, we must confine ourselves primarily to reactions involving C, N, and H. Of particular interest here are syntheses starting with HCN and HC_3N, both of which are known to be present on Titan. The question is whether or not the resulting polymers are of any interest to prebiotic chemistry. Adenine can be synthesized from hydrogen cyanide dissolved in ammonia. One can then ask whether this same synthesis can be achieved in the conditions existing in Titan's atmosphere or in the reservoirs of liquid hydrocarbons on the surface.

The time scale available becomes very important. The evolution of complex chemical systems over very long periods of time is difficult to model because of the uncertainties—no matter how small—about the exact reaction mechanisms and the rate constants. Thus the lakes or oceans on Titan may contain not only dissolved HCN but also some NH_3 formed from the irradiation of N_2 and H_2 by cosmic rays. With almost no time limitation, the formation of adenine and other purines on Titan cannot be excluded in spite of the low surface temperatures.

All of the purines and pyrimidines found in DNA and RNA have been identified in organic matter in meteorites, but how they were formed remains a mystery. The discovery of such compounds—or others of equal or greater complexity—on the surface of Titan would provide valuable insights into the nature of low-rate chemical and physical processes for the formation of organic compounds in a well-defined natural environment. These insights could then be used in conjunction with laboratory experiments to understand the prebiological chemistry on the primitive Earth. The changes in atmospheric composition on Titan over geologic time and the possibility that reaction products from ancient times are well-preserved on Titan's surface adds to the exobiological interest in future investigation of this giant satellite.

The Next Steps

To make real progress from our present position we need answers to some fundamental questions that have been raised by the observations available to date. Is that hypothetical ocean really there? What is its composition, depth, and lateral extent? What landforms are present and to what extent are they coated with precipitated aerosols? What is the composition of these aerosols and how exactly are they produced?

This last question leads us directly into the problem of chemical evolution in the atmosphere. We are looking here for evidence of any preferred pathways from nitrogen, methane, and carbon monoxide to more complex substances. To define such pathways and to understand the atmospheric chemistry we also need to know what other trace constituents are present and what sources of energy are available at different altitudes and latitudes. What causes the evident differences between the chemistry occurring at the poles and near the equator? What is responsible for the hemispheric difference in the reflectivity of the aerosol layer? Is there additional chemical processing at the surface driven by GCR or escaping internal heat? What can the noble gas abundances and their isotopic ratios tell us about the origin and evolution of Titan's atmosphere?

This list of questions can be extended for several pages, but the items already cited are sufficient to show the need for a new mission that returns to the Saturn system to make additional measurements. This mission should include a Saturn orbiter and a Titan probe. The orbiter would carry a radar mapper to define the satellite's surface topography and an array of infrared and microwave spectrometers to determine atmospheric composition and its variations with altitude and latitude. The probe would have a GCMS (Gas Chromatograph Mass Spectrometer) for studying gas composition connected to an aerosol pyrolyzer to analyze solid and/or liquid materials. A laser diode spectrometer could provide additional chemical analysis while defining the altitudes and scattering properties of clouds and hazes. A radar altimeter and a descent imager would furnish information on the nature of the surface during the probe's descent while, after landing, the GCMS could provide additional information on surface composition.

Both the probe and the orbiter would carry a variety of additional experiments as well, but the complement described above would already give us the basic information we seek. Just such a mission, called Cassini after the first director of the Paris Observatory who made many discoveries through his studies of the Saturn system, is currently planned as a joint enterprise between NASA and the European Space Agency. The Cassini Mission, with its Huygens probe into Titan's atmosphere, is scheduled for a late 1996 or early 1997 launch.

While we are awaiting the Cassini Mission there are still more observations that can be made from Earth. Among these, we point out the opportunity over the next decade for radar studies of Titan with the Arecibo and Goldstone antennas which should be able to determine whether or not a global ocean is present. There are also a number of additional infrared, submillimeter, and passive microwave observations that can be made as the sensitivity of detectors and the size of available telescopes increases during the next few years.

But there is simply no substitute for being there, so we look forward to the prospect of results from the Cassini Mission early in the next century.

Conclusions

The importance of Titan goes far beyond the exobiological aspects we have emphasized here. This satellite may be viewed as a member of a class of icy-rocky objects that includes comet nuclei and the planetesimals that formed the cores of the giant planets as well as the icy satellites that surround them. Studying the volatiles on Titan, trying to understand their origin, and investigating the subsequent chemical reactions they undergo will provide insights into a variety of other solar system objects as well.

Nevertheless, Titan's significance for exobiology may be its most outstanding characteristic. The opportunity to investigate the results of contemporary chemical reactions in a reducing atmosphere under totally natural conditions is a little like the anthropologist's dream of studying a primitive tribe that has been kept completely out of contact with any advanced civilization. It is even difficult to know in advance exactly what we will learn, and that too is part of the attraction this shrouded world presents.

Additional Reading

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Cosmic Dust

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D. E. Brownlee and
S. A. Sandford

Dust is a ubiquitous component of our galaxy and the solar system. Its average spatial density is less than one particle per cubic kilometer but the total number of particles is large. The particles are so prevalent that both interstellar and interplanetary dust can readily be seen with the naked eye under dark sky conditions. Zodiacal light, a glow usually seen in the east before dawn twilight and in the west following sunset, is sunlight reflected off dust particles orbiting the Sun. The dark band marking the central plane of the Milky Way is caused by absorption of background starlight by dust concentrated in the plane of our galaxy. In fact, dust is so prominent that light typically travels only a thousand years in the plane of our galaxy before it is absorbed by dust. The distance traversed in this time is only 1% of the diameter of the galaxy.

Interstellar dust is the predominant form of the condensable elements in the galaxy that are not in stars. The grains form in gas outflows from stars and they are processed, perhaps even destroyed and reformed, in the interstellar medium and molecular clouds. Interplanetary dust is debris recently liberated from comets and asteroids within the solar system. There is a strong link between interplanetary and interstellar dust. Prior to

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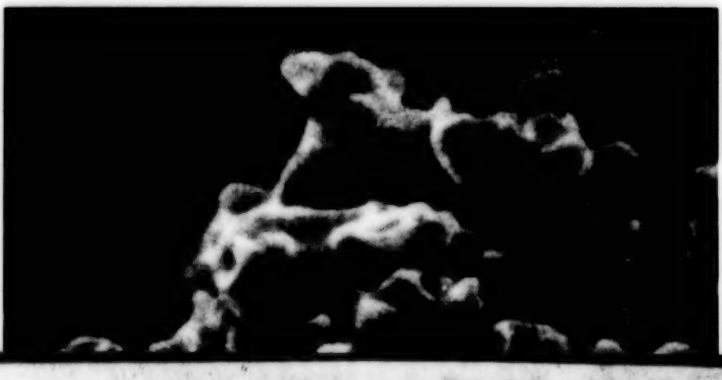


formation of the solar nebula, most of the atoms heavier than helium that found their way into our solar system were contained in interstellar grains. Some of these grains were incorporated into comets and asteroids. Comets forming in the outer fringes of the nebula presumably contain relatively higher abundances of presolar grains. Many of the grains preserved in comets may then be older than the Sun and planets. Comets are thus vehicles potentially capable of carrying the products of interstellar chemical processes directly to planets. They are believed to contain substantial organic components produced in space by nonequilibrium reactions such as catalysis and radiation processing of condensed volatiles.

The collection and analysis of extraterrestrial dust particles is important to exobiology because it provides information about the sources of biogenically significant elements and compounds that accumulated in distant regions of the solar nebula and that were later accreted onto the planets. Both interstellar and interplanetary dust seeded the early Earth with elements and compounds that may have played important roles in the origin and development of life. Interplanetary dust particles (IDPs) that are collected in the stratosphere and analyzed in the laboratory are preserved fragments from asteroids and comets, bodies which formed in the outer regions of the solar nebula, are relatively rich in volatiles, and which are a source of abiotic organic molecules that have fallen to Earth throughout its history.

Interplanetary dust particles accreted by the Earth must first enter the atmosphere. They enter the atmosphere at velocities ranging from 11.2 km/s, the Earth's escape velocity, to a maximum of 72 km/s, the velocity of a head-on impact with a particle in a retrograde parabolic orbit. The kinetic energies corresponding to these velocities are capable of melting or vaporizing the incoming particles if the energies are converted to internal heat. In the case of conventional meteorites, the surfaces are heated to vaporization temperatures, although the short time scale of heating and the combined effects of thermal inertia and conductivity confine melting and strong heating to the outer millimeter or so of the body. The interiors of the meteorites remain essentially unheated. Dust particles in the 0.1 mm to several mm size range typically also experience melting from atmospheric drag but, because of their smaller size, they tend to have isothermal interiors and they melt completely, resulting in the loss of volatiles such as sulfur, sodium, and organic matter. Such particles form melt spheres (cosmic spherules) that have been collected in abundance from the ocean floor and polar ice deposits.

Particles smaller than 0.1 mm, however, enjoy a special advantage over conventional meteorites and larger dust particles. These grains can usually enter the atmosphere without melting. This is possible because the particles only travel at high velocity at altitudes above 90 km where the air is very tenuous. For a given velocity, the frictional power density generated on the face of a particle depends only on the local air density. At 90 km, the frictional energy is generated slowly enough that it can be thermally reradiated without the particle being heated to its melting point. Smaller particles decelerate at higher altitudes and for a given initial velocity there is a limiting size below which no melting will occur. Particles that do not melt are called micrometeorites and are genuine samples of interplanetary dust. Most particles below 50 µm in size do not melt upon atmospheric entry and the survival of solar flare cosmic ray tracks and minerals with low thermal stabilities in the particles indicate that typical 10 µm particles are not heated above 600°C.

An additional factor favoring the survival of interplanetary dust during atmospheric entry is the fact that they are not subjected to large ram pressures during atmospheric deceleration. The ram pressure experienced by dust particles is orders of magnitude smaller than must be survived by larger objects such as conventional meteorites. This allows fragile materials to enter as dust but prevents them from surviving as millimeter or larger sized objects. At a given velocity the ram pressure is proportional to the ambient air density. The maximum ram pressure experienced by a micrometeorite that decelerates near 90 km is 100,000 times smaller than that experienced by a 10-kg rock that has retained its cosmic velocity down to an altitude of 40 km. Even the most fragile conventional meteorite is a factor of 100 times stronger than the materials observed in cometary meteor showers. The altitudes and velocities where cometary meteors are observed to fragment imply typical crushing strengths of 10^5 dynes/cm^2 and in the extreme case of the Draconid meteors (the youngest meteor stream) the pressure is only 10^3 dynes/cm^2. Micrometeorites this fragile can enter the atmosphere without fragmentation whereas larger objects cannot.

Thus, typical cometary rocks cannot enter the atmosphere without being crushed. Once crushed, most fragments should either melt or vaporize because fragmentation occurs at lower altitudes where frictional heating of small hypervelocity particles is severe. If comets do not contain any strong centimeter and larger components, then dust would be the only meteoroid type that could carry cometary molecules to the Earth's surface.

Micrometeorites ranging in size from 5 to 50 µm are routinely collected in the stratosphere using high-altitude aircraft. The particles are characterized and curated for distribution to investigators worldwide. Most of the particles are easily identified as extraterrestrial on the basis of their elemental composition which is similar to that of primitive meteorites but quite different from most terrestrial particles. Typical particles are composed of Mg, Si, Fe, S, Ca, Na, Al, Ni, Mn, Cr, and Ti in ratios that match their relative abundance in the Sun. Oxygen and carbon are also major constituents but, as in meteorites, their relative abundances are less than solar. An extraterrestrial origin can be proven for any given particle by detection of He implanted by the solar wind or tracks in mineral grains produced by solar flare cosmic rays. As previously mentioned, larger particles are also recovered from deep sea sediments and from polar ice (Greenland and Antarctica). Particles larger than 0.1 mm are usually melted, but a few particles up to 1 mm survive without any melting. These rare, giant micrometeorites are presumably particles that entered at relatively low velocity and very low incident angles. In low-angle impacts the hypervelocity interaction with the atmosphere occurs at higher altitudes and results in prolonged but less severe heating.

General Properties of Interplanetary Dust

The most common IDPs collected from the stratosphere are black, fine-grained materials that have major and minor element compositions similar to those of carbonaceous chondrites, i.e., primitive meteorites. Particles with this composition are referred to as being "chondritic" even though they may have no actual relation to chondritic meteorites. Some micrometeorites do not have chondritic compositions, but many of these are large single-mineral grains that have surficial debris indicating that they were previously embedded in larger samples of fine-grained chondritic material. The "chondritic" particles are small (typically less than 20 μm in diameter) and classifying them into meaningful groups is much more difficult than with conventional kilogram-sized meteorites. The primary properties that can be readily measured are morphology, elemental composition, and mineralogical composition as determined by both electron microscopy and infrared spectroscopy. The particles can be grouped into two major classes: those dominated by hydrated (layer-lattice) silicates and those dominated by anhydrous silicates, of which pyroxene and olivine are the most common. These differences are clearly seen in infrared spectra of individual particles (fig. 8-1).

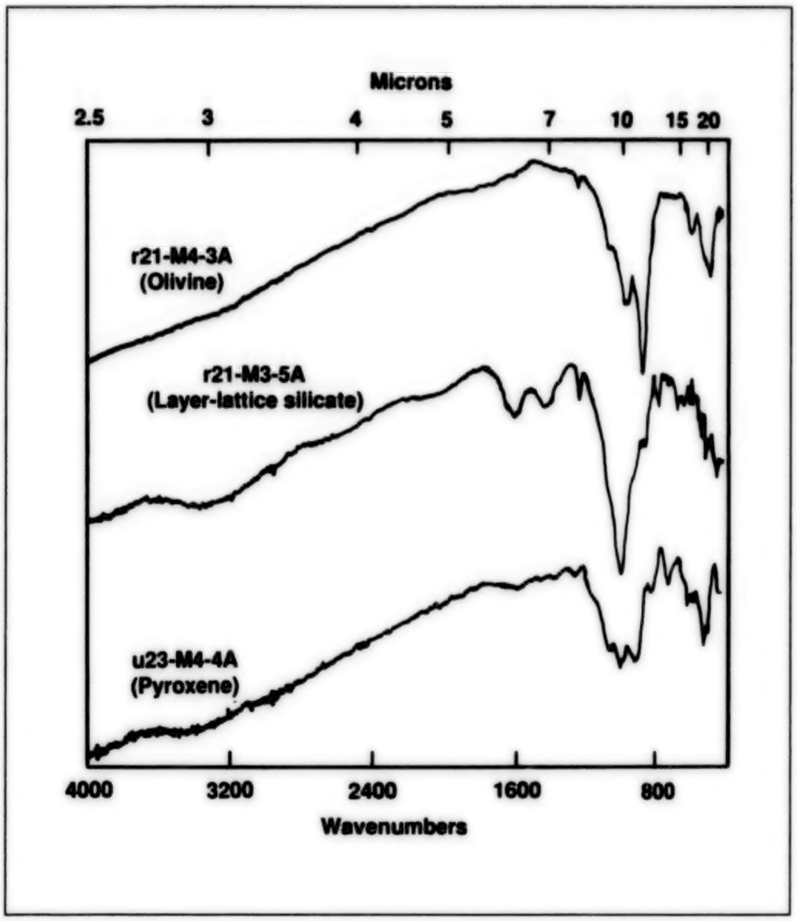
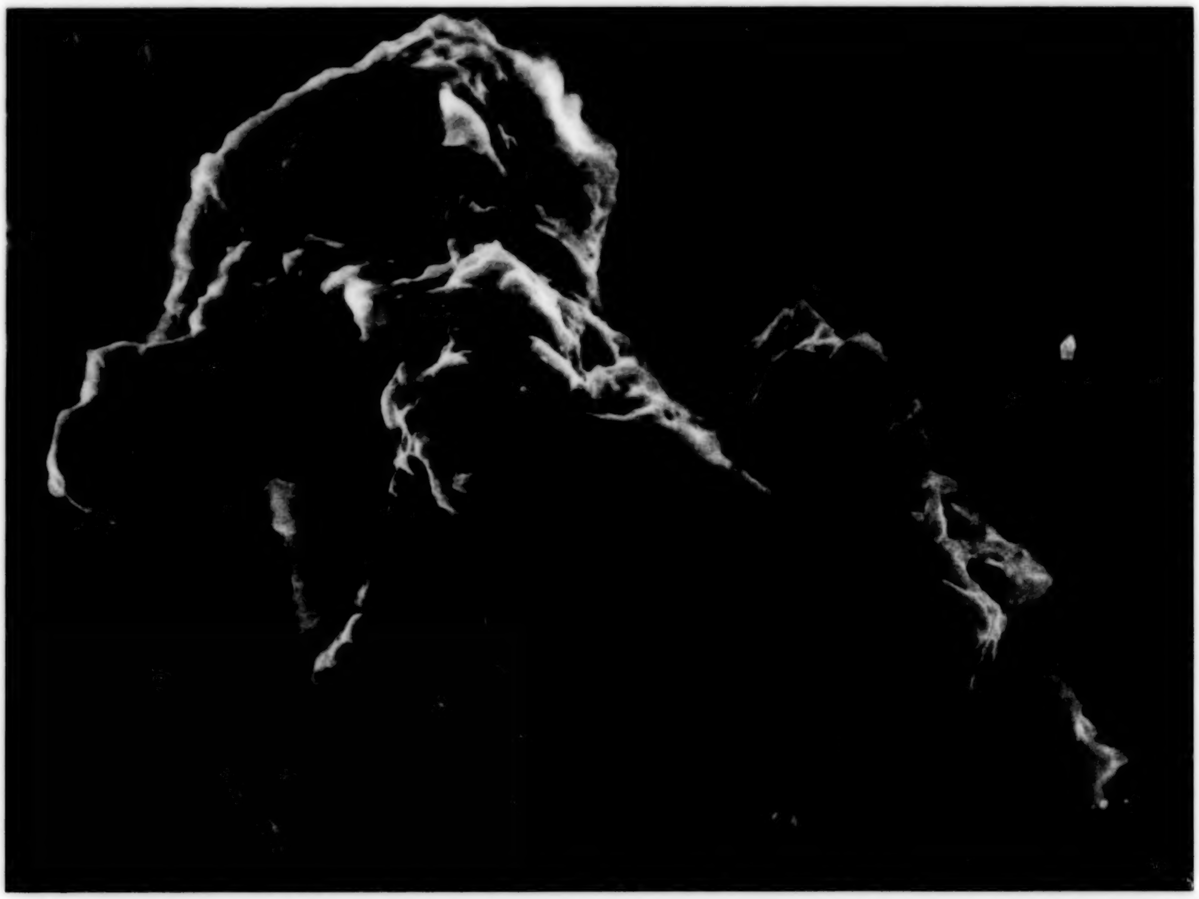


Figure 8-1. The three major infrared spectral IDP classes. From top to bottom, these spectra are representative of IDPs dominated by the minerals olivine, layer-lattice silicates, and pyroxenes, respectively. The spectra can be easily separated by the profiles of their characteristic 10 μm (1000 cm^{-1}) Si-O stretching features. The spectra also show longer wavelength features due to Si-O-Si bending vibrations in the silicates. In addition to the Si-O bands, spectra from the particles dominated by layer-lattice silicates also show bands centered near 3.0 and 6.0 μm (3330 and 1670 cm^{-1}) due to O-H stretching and H-O-H bending vibrations in adsorbed and absorbed water, and bands near 6.8 and 11.4 μm (1470 and 875 cm^{-1}) due to C-O stretching and CO$_3$ scissoring vibrations in carbonates. Note the minor features at 3.37, 7.94, and 12.53 μm (2970, 1260, and 798 cm^{-1}) due to silicon oil in the middle spectrum.

The hydrated particles are similar to two types of primitive meteorites known as the CI and CM carbonaceous chondrites. These meteorites are dominated by layer-lattice silicates and they are the only chondrites that have carbon abundances appreciably above 1% by weight. The bulk of hydrated IDPs is composed of tangled masses of layer-lattice silicate sheets and fibers. There are two subgroups of hydrous IDPs, one dominated by serpentine-like minerals with 7 Å basal spacings typical of CI and CM meteorites, and one dominated by a layer-lattice silicate with 10-12 Å spacings not seen in most meteorites. Like the conventional meteorites, most of the hydrated IDPs are compact objects having only minimal pore spaces (fig. 8-2). Like the CI and CM chondrites, many of these particles have properties consistent with processing by aqueous alteration on a parentbody. These include the existence of carbonates, clusters (framboids) of rounded magnetite grains, and redistribution of Ca and Mg, elements that often are components of water-soluble

Figure 8-2. A scanning electron microscope photograph of a particle dominated by hydrated (layer-lattice) silicates. These particles are generally compact and contain little void space. The white scale bar on the photograph is 10 μm long.



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minerals. The CI and CM chondrites are the most primitive meteorites known on the basis of match with the elemental composition of the Sun, but they are heavily altered rocks that have been processed in a relatively warm and wet parentbody. Their parentbodies were almost certainly asteroids as it is difficult to imagine aqueous alteration occurring on comets, small bodies which sublime when warmed above cryogenic temperature. By analogy, it is likely that many and possibly all of the hydrated IDPs are also of asteroidal origin.

The second general class of IDPs is dominated by anhydrous minerals such as olivines, pyroxenes, and iron sulfides. These particles are similar to CI and CM chondrites in bulk elemental composition, but in terms of mineralogy and structure they are unique and have no close analogs among meteorites. These IDPs should be considered a new type of carbonaceous chondrite that apparently does not survive atmospheric entry in the form of conventional, large meteorites. They are carbon-rich and composed of roughly equidimensional grains ranging in size from micrometers to less than 100 Å. They also contain amorphous materials such as glass and disordered carbon. Some of the anhydrous particles are highly porous (as in the Frontispiece) and resemble gravel aggregates or clusters of grapes. The high porosity suggests a similarity to the porous, fragile materials observed in cometary meteor streams. High porosity and weak structure are evidence against compaction processes that probably occur in asteroids due to meteoroid impact and, in some cases, self-gravitation. The submicrometer and micrometer pore spaces in IDPs are likely to be voids produced by sublimination of volatile materials such as ice or perhaps organic compounds with high vapor pressures.

The Carbonaceous Component of IDPs

Carbon is a major element in IDPs, but it is difficult to study because of fundamental analytical limitations and ever present contamination problems. Studies of the carbon in IDPs are in an early stage of development and the nature of its form and distribution in this material is not yet well established. The carbon abundance in the particles is not well determined but it appears to be in the 5-15% range by weight for most particles. In terms of atom fraction this makes carbon the second most abundant element after oxygen in IDPs. This carbon content is intermediate between that of the most carbon-rich meteorites known and that measured in dust from comet Halley by Soviet and European spacecraft.

Electron microscope studies have shown that most of the carbon in IDPs is amorphous. It occurs in micrometer and submicrometer grains, as a component of submicrometer lumps composed of 100 Å mineral grains and carbon, and as 100 Å films on mineral grains. Oxygen and nitrogen do not appear to be major constituents of the carbonaceous matter and from an electron microscopy viewpoint the material looks like elemental carbon (although the techniques used could not detect hydrogen). The H, C, N, and O containing "CHON" grains identified as abundant micrometer-sized particles in the Halley coma have not been identified in IDP samples. It is possible that the cometary CHON grains consist of relatively volatile materials and they do not survive the prolonged exposure to space and subsequent entry into the Earth's atmosphere.

Part of the problem with studying the carbonaceous material in IDPs by electron microscopy is that this technique is most sensitive to crystalline phases, while most of the carbon in IDPs is amorphous. The most common carbon-bearing crystalline phases found in IDPs include Mg- and Ca-rich carbonates, iron carbides, lonsdaleite, and trace amounts of graphite. The iron carbide is ε-iron carbide and it is found in association with elemental carbon, Fe-Ni metal, and magnetite. The carbides are believed to have been formed by catalytic reaction of CO gas on grain surfaces. This most probably occurred in the solar nebula. Graphite is easy to identify in the electron microscope because of its distinctive 3.4 Å basal spacings. Yet graphite is rarely seen in IDPs. If the anhydrous particles are cometary, it is clear that graphitic carbon is not a major constituent of comets. If preserved grains are abundant components of comets, then it would also follow that graphite is not abundant in interstellar grains either.

Additional information about the form of the carbonaceous material in IDPs can be gleaned from their infrared transmission and Raman spectra. As was shown earlier (fig. 8-1), the infrared transmission spectra of individual IDPs are dominated by their abundant silicate minerals. However, some spectra also show the presence of a weak absorption feature near 3.4 μm (2940 cm^{-1}). This spectral position is characteristic of C-H stretching vibrations and suggests the presence of hydrocarbons. Unfortunately, the stratospheric IDPs are collected on impact collectors coated with a thin layer of silicone oil, which itself produces an absorption feature near 3.4 μm. Thus, it is difficult to ascertain what fraction of the observed 3.4 μm band in the IDP spectra is due to indigenous hydrocarbons and how much is due to residual silicone oil that was incompletely removed from the particle during curation. The layer-lattice silicate in figure 8-1 provides a good example of a particle in which the majority of the observed 3.4 μm feature is probably due to silicone oil. The presence of features at 7.94 and 12.53 μm (1260 and 798 cm^{-1}) are also characteristic of silicone oil. There are particles, however, in which a 3.4 μm feature is present and unaccompanied by the corresponding silicone oil features at 7.94 and 12.53 µm. In these cases, we may be seeing evidence of indigenous carbonaceous materials, although it has not been possible up to the present time to eliminate the possibility of other forms of contamination.

If the 3.4 µm features observed in IDP spectra are due to indigenous materials, their positions and profiles indicate that aliphatic compounds are present. The overall IDP feature, when seen, is similar to that observed in spectra taken toward the galactic center and from residues produced in laboratory experiments in which simple mixed molecular ices are irradiated with ultraviolet photons or ions. All three cases represent material which has had volatile components removed. In the case of the IDPs, the more volatile components have been removed by exposure to the Sun in interplanetary space and by heating during atmospheric entry. In the case of the galactic center, the volatiles have been removed by long duration exposure to the diffuse interstellar medium.

Unfortunately, our inability to rule out contamination by silicon oil or other lab materials makes uncertain any conclusions derived from the observed 3.4 µm IDP features. Further work will be needed before the infrared spectra can place any strong constraints on the composition of the carbonaceous compounds in IDPs.

Raman spectroscopy has been somewhat more successful in providing information about the state of the carbon in IDPs. As can be seen in figure 8-3, Raman spectroscopy is particularly sensitive to the CC vibrations within the aromatic molecular units in carbonaceous materials. The top five spectra, taken from laboratory standards, demonstrate how the Raman spectra of carbonaceous materials change as the material becomes less ordered and the size of the aromatic domains decreased. Graphite, a material in which the aromatic domains are very large, produces a single intense first-order band at 1581 Δcm^{-1} and a weaker second-order feature near 2700 Δcm^{-1}. As the aromatic domains decrease in size, both of these bands grow broader and new features appear near 1350 and 2950 Δcm^{-1}. The bottom two spectra were taken from individual IDPs. Comparisons of the widths and relative strengths of the features in both sets of spectra show that the carbonaceous material in IDPs contains aromatic domains which are generally smaller than 25 Å.

In addition to the vibrational features, many of the Raman spectra of IDPs also show a broad photoluminescence feature arising from electronic transitions (see, for example, the bottom spectrum in figure 8-3). The position and strength of the luminescence varies from particle to particle, but typically the short wavelength limit of the feature is around 5400 Å and the feature peaks between 5900 and 6400 Å. This red luminescence is of particular interest since it has been known for some time that small bodies in the outer solar system show a general tendency to grow darker and redder with heliocentric distance. Photoluminescence associated with carbonaceous materials similar to those in the IDPs may contribute to the observed reddening.

It is also interesting to note that, while the particles are dominated by silicate minerals (as evidenced by the infrared spectra and the electron microscope studies),

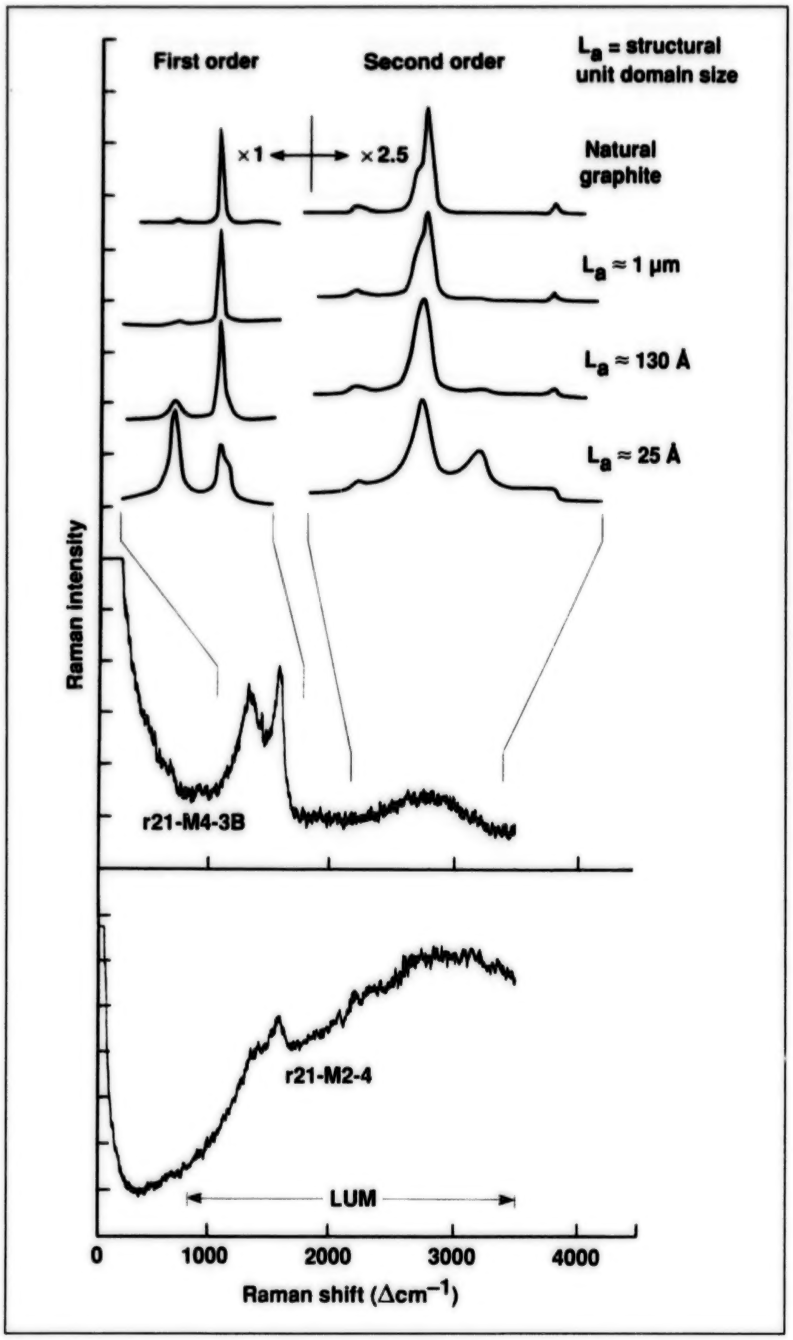


Figure 8-3. Raman spectra of several laboratory standards and IDPs. The top five spectra were taken from laboratory materials having decreasing degrees of crystalline order. The bottom two spectra were taken from individual IDPs. Comparison between the two sets of spectra show that the aromatic domains in IDPs are mostly smaller than 25 Å. Note the broad photoluminescence apparent in the lowest IDP spectrum.

the Raman spectra seldom contain bands due to these minerals. This suggests that the carbonaceous material in IDPs effectively "screens" the silicates from visible photons. This explains how the IDPs can have relatively dark appearances and yet still be dominated by silicate minerals. This observation has implications for the many observations obtained during comet Halley's recent apparition.

Many of the collected IDPs are likely to be derived from comets. Figure 8-4 shows comparison between a 5-13 µm (2000-770 cm^{-1}) spectrum of comet Halley and a composite spectrum derived from IDP data. The IDP composite spectrum consists of a mixture of roughly 65% olivine-rich, 35% pyroxene-rich, and 10% layer-lattice silicate-rich IDPs. Note that not only is the 10 µm (1000 cm^{-1}) silicate feature reasonably well fit, but the Halley spectrum also shows a feature near 6.8 µm (1470 cm^{-1}) which is consistent with the presence of

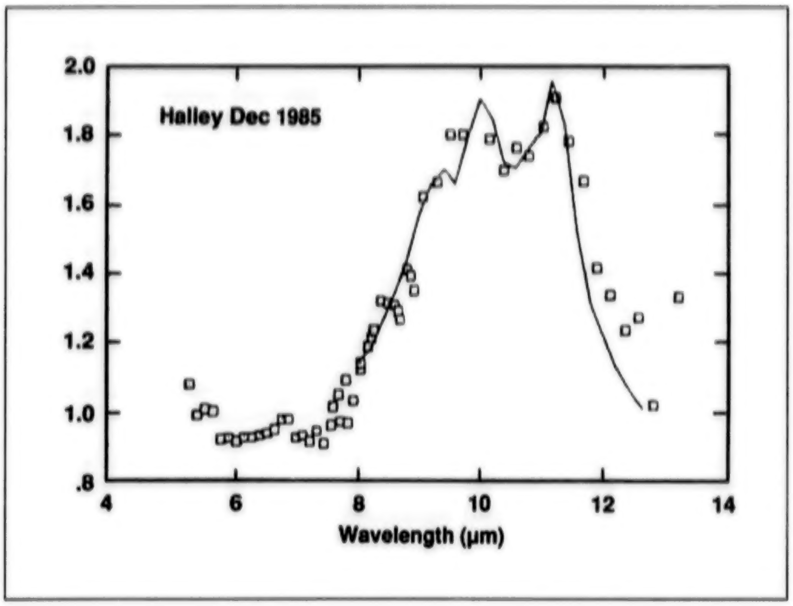


Figure 8-4. A comparison between the spectrum of comet Halley (points) and a composite spectrum of several IDPs (solid line). The IDP composite contains contributions for all three IDP infrared classes in the relative proportions of 55% olivines, 35% pyroxenes, and 10% layer-lattice silicates.

carbonates in the abundances they are normally found in the IDPs dominated by layer-lattice silicates. Similar fits to the IDP data have also been found for comets Wilson and Giocobini-Zinner. The match between the IDP spectra and the cometary spectrum and the detailed structure within the cometary 10 µm feature demonstrate that these comets contain large amounts of silicates and that these silicates are crystalline. Thus, analogous to the collected IDPs, the low observed albedo of Halley does not necessarily imply the comet is dominated by carbonaceous materials. Instead, the abundant silicates may be screened from visible photons by less abundant carbonaceous materials.

There is some indication, however, that the carbonaceous component in comets includes a component that is not present in the collected dust. Infrared spectra of comets Halley and Wilson have been shown to contain emission features near 3.4 µm (2940 cm^{-1}) that are diagnostic of C-H stretching vibrations in hydrocarbons. These features fall at shorter wavelengths than similar features observed in IDPs and toward the galactic center. As the collected dust and the dust toward the galactic center has undergone some thermal processing, this suggests that comets contain some carbonaceous materials that have relatively low volatilities. This material may remain with the dust long enough to be observed in the telescopic data but not survive long enough to be observed in the collected interplanetary dust.

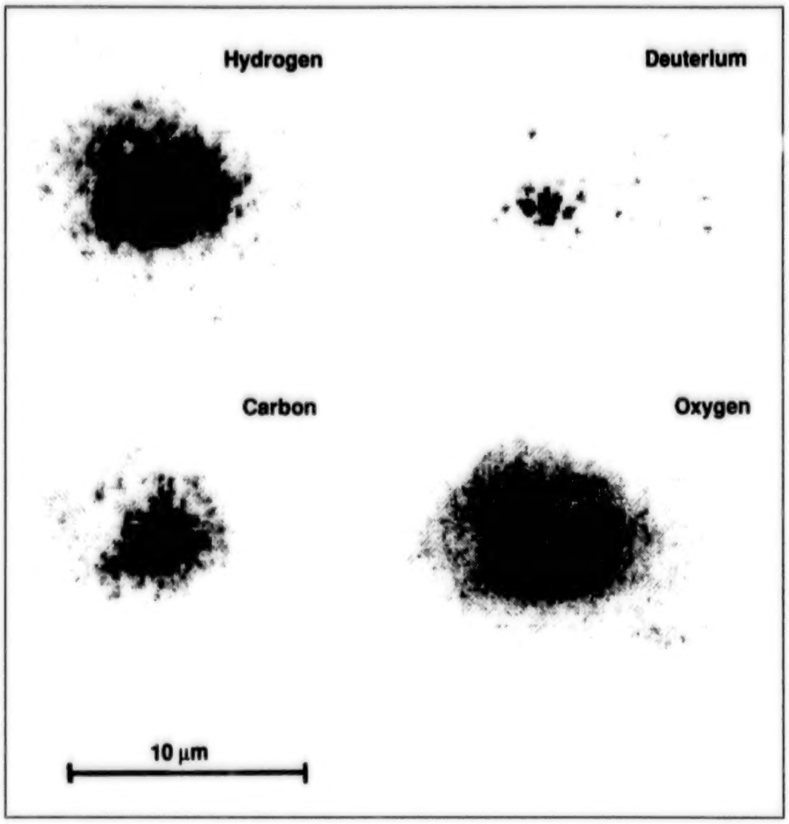
In summary, while the presently available information about the chemical state of the carbonaceous material in IDPs is quite limited, there is presently nothing inconsistent with the view that the majority of the carbonaceous material in IDPs is similar to the kerogen-like materials in carbonaceous meteorites, i.e., it consists of a network of small aromatic domains randomly interlinked by more-aliphatic bridges. This material contains minor amounts of O and N and resides in the particles both as separate clumps and as matrix material.

The Presence of an Interstellar Component

The actual formation sites of the carbonaceous material in IDPs are unknown. Certainly, some of the material must have formed in the early solar nebula. However, given that much of the dust may come from comets and that comets are likely to be relatively primitive bodies, the intriguing possibility exists that a fraction of the material has an interstellar origin. This possibility is of special interest to biogenic studies since the interstellar medium offers a rich variety of environments in which carbon chemistry can occur. In addition to normal gas phase chemistry, carbon-containing compounds can form in the interstellar medium via ion-molecule reactions, reactions between gas phase molecules catalyzed on grain surfaces, and reactions in ice mantles induced by ultraviolet photons and cosmic rays. This wide variety of chemical processes potentially provides for the production of a rich assemblage of biogenically interesting compounds. Given the diversity of chemical processes operating in the interstellar medium, it is worth asking whether there is any evidence that the carbonaceous material in IDPs contains an interstellar component. The answer, provided by isotopic studies, is yes!

Many of the collected IDPs are found to contain large deuterium enrichments which are inconsistent with an origin in the solar system. The spatial distribution of the carrier of the deuterium excess in IDPs is not uniform, but instead seems to be confined to "hot spots" whose dimensions are smaller than a micrometer (fig. 8-5). This suggests that the carrier of the deuterium consists of small "grains" which are distributed randomly within

Figure 8-5. "Pictures" of an IDP taken at different masses using an ion microprobe. Note that while the elements oxygen, carbon, and hydrogen are distributed more or less uniformly throughout the particle, the deuterium is concentrated in a localized region. The size of the observed deuterium-rich "hot spot" (approximately 1 μm in diameter) is on the order of the spatial resolution of the ion probe. As a consequence, only a lower limit to the deuterium enrichment in this spot can be obtained. The D/H ratio in the hot spot is found to be greater than a factor of 10 times the terrestrial value, making this the most deuterium-rich naturally occurring sample ever examined in the laboratory. The scale bar is 10 μm long. (Picture courtesy of R. Walker, Washington University at St. Louis).



the overall particles. Unfortunately, because of the small size of the carrier, it is not yet clear whether the deuterium-rich hot spots represent true, unaltered interstellar grains that have been incorporated into the larger IDP, or whether they represent material partially altered in the solar nebula which still has a "molecular memory" of such a grain. The deuterium enrichments seem to correlate loosely with carbon concentration, suggesting a carbonaceous carrier phase is responsible. However, the carbon isotopic system is clearly decoupled from the deuterium enrichments. The D/H ratio is not found to correlate with the $^{13}C/^{12}C$ ratio in the same particles, nor does the $^{13}C/^{12}C$ ratio show the same heterogeneity exhibited by the D-H system. This implies that, while the deuterium is carried by a carbonaceous phase, the carrier and the enrichment need not have formed in the same place, at the same time, or by the same process(es).

The chemical origin of the deuterium excess is uncertain. Deuterium enrichment via chemical reactions requires very low temperatures ($T < 100$ K), but at these temperatures the reaction rate is extremely slow. The reaction threshold is greatly reduced, however, if one of the reactants is an ion. For this reason, it is felt that ion-molecule reactions taking place in dense, cold interstellar clouds may be the source of the observed deuterium enrichments. An additional means by which the deuterium enrichment could have occurred in the interstellar medium involves selective photodissociation reactions. Interactions of ultraviolet photons with aromatic molecules in the interstellar medium should result in the photodissociation of some of the molecules' peripheral hydrogen atoms. Since the zero point energy of the C-D bond in polycyclic aromatic hydrocarbons is lower than that of the C-H bond, deuterium is less likely to be dissociated. The dissociation site is then free to capture a new atom. Over an extended period of successive photodissociation and recapture, a deuterium enrichment of the molecular aromatic population is expected. In any event, whether the observed enrichments are produced via ion-molecule reactions or selective photodissociation, their presence in IDPs clearly points to material having an interstellar origin.

It should be noted, however, that while the presence of an exotic isotopic component can prove that interstellar material is present, the lack of an exotic isotopic signature does not necessarily imply formation in the solar system. Since the solar system is presumably derived from more or less normal interstellar material, we might expect most interstellar material to have "normal," i.e., solar system-like isotopic abundances. Thus, the interstellar material evidenced by the deuterium enrichments represents a lower limit to the abundance of interstellar material present.

Summary

Collected samples of interplanetary dust are of exobiological interest since these particles contain a variety of carbon-bearing compounds and offer an efficient means of seeding planets in the early solar system with these materials. Because of their small size, IDPs can be decelerated from cosmic velocities at high altitudes in planetary atmospheres. As a result, much of the dust is captured without experiencing thermal excursions sufficient to melt or vaporize them. Any carbonaceous materials indigenous to the particles are then deposited on the surface of the planet.

While several minor carbon-bearing phases have been identified in IDPs, the majority of the carbonaceous material in IDPs seems to consist of a disordered material rich in C and H and containing minor amounts of

O and N. Spectroscopic evidence suggests that this material contains aromatic molecular units smaller than 25 Å in size and possibly aliphatic hydrocarbons as well. In so far as a comparison is possible, the majority of the carbonaceous material in IDPs appears to be similar to the kerogen-like material found in primitive carbonaceous meteorites, a material consisting of small aromatic moieties that are randomly interlinked by aliphatic structures.

The collected dust can provide reasonable matches to astronomical spectra of comets. The known extraterrestrial nature of these particles and their spectral similarity to primitive astronomical objects all argue that the collected IDPs contain "primitive" materials that may have been present during the formation of the solar system.

Isotopic studies of the IDPs demonstrate that many of them contain small (<1 micron) components that are greatly enriched in deuterium. The magnitudes of the observed enrichments are not explainable in terms of solar system processes and indicate the presence of interstellar material. It is not clear at this time whether the carriers of the isotopic anomalies represent true, unaltered interstellar dust grains, or whether they represent an altered component with a molecular "memory" of original interstellar grains. In any event, it is clear that the collected IDPs contain molecular material not only from the early solar system, but also from the interstellar medium. Since the interstellar medium contains a wide variety of environments in which many different chemical processes can occur, it would not be surprising if IDPs contain a varied assemblage of compounds of exobiological interest.

Finally, we note that there are a number of proposed spacecraft missions that could provide important new insights into the nature of interplanetary dust and comets. The most spectacular of these is the Rosetta Comet Sample Return Mission planned by the European Space Agency. This Mission will be flown in the next century and is designed to land on a short period comet, collect a core sample, and return the sample to Earth while retaining cryogenic temperatures. A nearer term mission is the Comet Rendezvous Asteroid Flyby (CRAF) Mission which is tentatively planned for a 1996 launch that will result in matching orbits with a short period comet. The spacecraft will be in close proximity to the comet for over a year and will measure the composition of the gases and dust emitted by the comet over much of its orbital cycle.

Additional Reading

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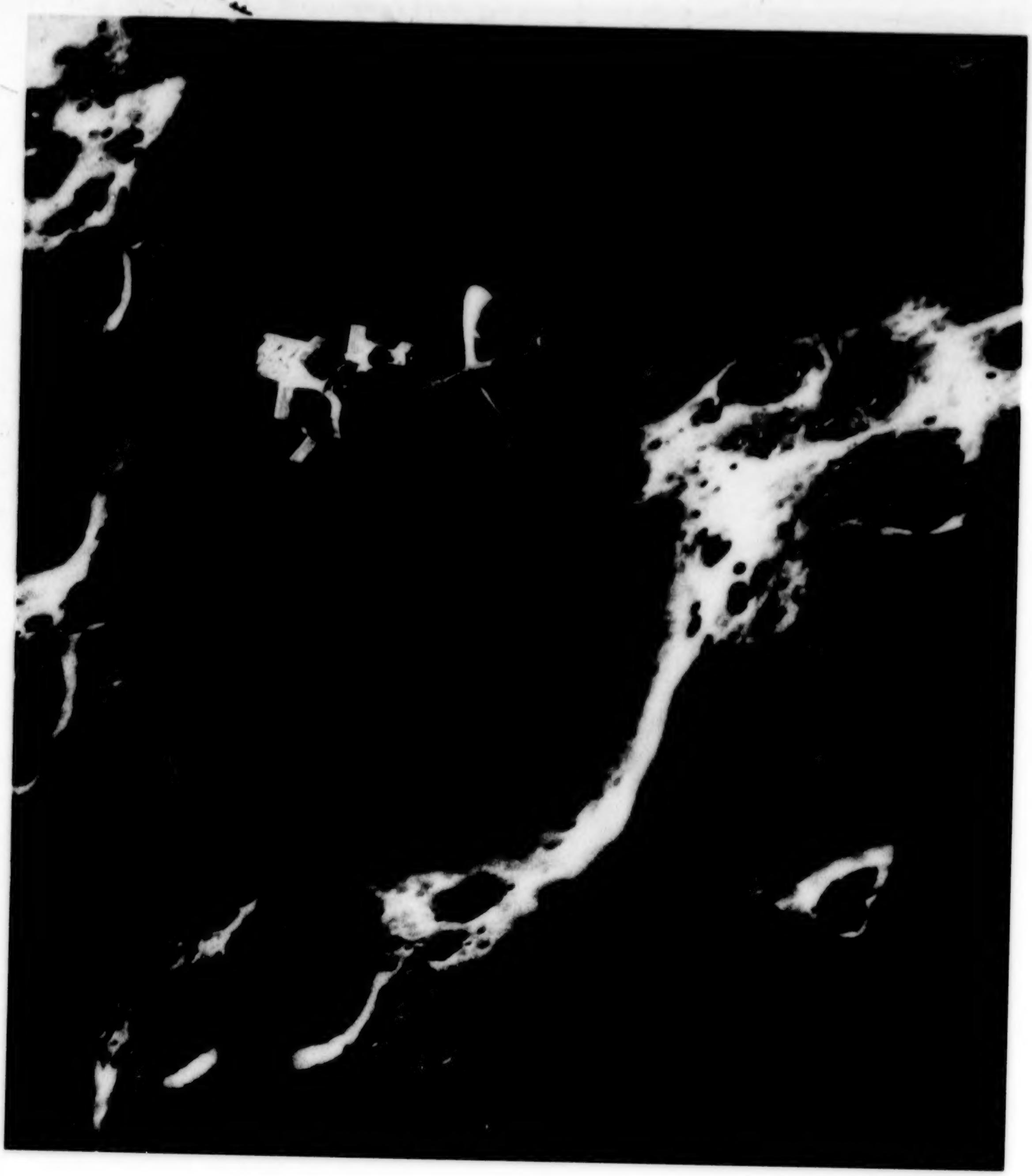

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Organic Material: Asteroids, Meteorites, and Planetary Satellites

N93-18555

D. P. Cruikshank and
J. F. Kerridge

Telescopic observations in *in situ* spacecraft investigations over the last two decades have shown that many planetary satellites, asteroids, and comets have surfaces containing very dark material that is either neutral (black) or red in color. Awareness of the presence of this very dark material probably originated with the 1971 discovery of the very low geometric albedo (0.03) of asteroid 324 Bamberga, determined from its easily detectable thermal radiation. Further infrared observations showed that many other asteroids in the main belt have low geometric albedos, and in 1975 it was established that dark asteroids are very common, particularly in the outer parts of the main belt. That the Trojan asteroids and the outer satellites of Jupiter are similarly very dark was first shown by infrared observations in 1977.

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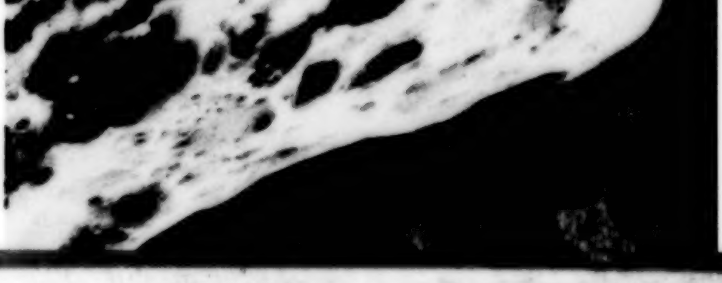


Most of the smallest planetary satellites, including the majority of the objects accompanying Mars, Jupiter, and Uranus, have surfaces of very low albedo, as revealed by spacecraft encounters and continued telescopic observations. The outer Jovian satellites and the two satellites of Mars are comparable in size and have low albedos and colors that are similar to the C-type (dark and neutral) asteroids. The small, inner satellites of Uranus discovered by Voyager are also low in albedo. The small, inner satellites of Saturn are an exception to the above examples because they lie in the range of geometric albedos from about 0.4 to 1.1. Phoebe, the outermost known satellite, has an albedo of 0.06 and therefore resembles the asteroids rather than the ice-covered satellites of the Saturnian system.

Saturn's large satellite, Iapetus, is an extraordinary object having a bimodal surface distribution of high and low albedo material. The hemisphere of Iapetus centered on the apex of its orbital motion has a very low albedo of 0.04 and is red, while the trailing hemisphere of the satellite has an albedo of about 0.5 and is composed of water ice.

Although comets are not the focus of this paper, the possible relationship of comets to asteroids, meteorites, and interplanetary dust (see below) requires that we discuss them briefly in the context of their dark-matter component. Evidence has been steadily accumulating that at least some, if not all, comet nuclei have surface deposits of low-albedo material. Telescopic observations of comet Halley on its approach to perihelion gave indications that its nucleus has a dark red surface of low albedo. The direct images of the nucleus of this comet by the Giotto and two Vega spacecraft unambiguously confirmed the pre-encounter deduction, showing that the nucleus has albedo 0.04-0.05.

Comet Halley contains high-molecular-weight organic molecules that have been detected both by direct sampling at the comet by the Vega and Giotto spacecraft and from ground-based telescopes. While the identification of the specific organics is difficult, the presence of aromatics, aliphatics, and amorphous heavy molecules can be reasonably concluded. Furthermore, the presence of these heavy organics, which often tend to be black, is entirely consistent with the low albedo of the comet nucleus derived from direct observations. Indeed, the formation of a nonvolatile crust of low albedo from cometary ices appears to be an inescapable consequence of ion irradiation of the outer few meters of the nucleus.

Beyond the direct observations of the nucleus of comet Halley, the accumulating telescopic data on other comets are giving a similar picture of the widespread occurrence of dark nuclei. Many periodic comets have nuclear albedos of 0.10 or less, with the majority studied so far, less than 0.07.

Except for Iapetus, the objects noted so far have relatively uniform surface properties of color and albedo. There are several other planetary satellites, and probably asteroids, with surfaces that include some fraction of dark material. For example, the five large satellites of Uranus have reflectance spectra indicating the dominance of water ice on their surfaces, but their albedos are too low for water ice, ranging from Umbriel at about 0.17 to Ariel at about 0.39. The dark material mixed with the water ice of the Uranian satellites is a neutral color, but its albedo remains unknown.

Pluto has a nonuniform brightness distribution across its surface, with a wide, circumplanetary dark band of relatively low albedo at the equator. Models of Pluto's surface constrained by the rotational lightcurve and the secular change in brightness due to changing observing aspect predict large surface areas of lower and higher albedo, as well as two polar caps. Charon, the satellite of Pluto, has an albedo of 0.4 (compared to Pluto's albedo of 0.6, both in the blue spectral region). We return to these two objects below.

It appears that dark matter occurs throughout the outer solar system, particularly on small bodies, but also mixed with other surface components on larger ones. At the present time, we know only that this low-albedo material can be reddish in color (reflectance increasing at longer wavelengths) or neutral. This simplistic observation may well belie a tremendous diversity in molecular or mineral composition.

Cosmic material of low albedo that is available to us for study on Earth consists of meteorites of certain classes and interplanetary dust particles collected from the stratosphere, the Greenland ice sheets, and the sea floor. It is natural to consider the possible connections of these materials to the comets and the asteroids, as mechanisms for their delivery to the Earth have been established and can be readily understood. Interplanetary dust particles are discussed elsewhere in this volume, but here we discuss the carbonaceous meteorites in the context of their organic content.

Carbonaceous Chondrites

Meteoritic organic matter occurs primarily in those meteorites termed carbonaceous chondrites that have the additional property that their chemical compositions are similar to that of the Sun. The two properties are related at least to the extent that preservation of organic molecules would have been aided by a history that led to minimal chemical fractionation. Nonetheless, it is important to note that even carbonaceous chondrites have not been immune to geochemical processing during their evolution and that the episodes of localized heating, impact brecciation, irradiation by solar flares and galactic cosmic rays, and aqueous mineralization recorded in them may well have altered their organic components and may even have been implicated in the origin of those components.

The molecular and isotopic composition of meteoritic organic matter has been the subject of numerous studies, but considerable work remains to be done, particularly by application of recently developed analytical techniques to issues that were studied at lower sensitivity and resolution in the 1960s and early 1970s. The principal objective of such studies is to discriminate between the different hypothetical origins that have been proposed for the organic matter. Variables used for that purpose have generally involved the molecular and isotopic distributions within specific classes of compounds, e.g., as a function of C number within homologous series. Patterns observed so far are difficult to fit within a single mode of synthesis and suggest instead multiple origins for the organic species now found in meteorites.

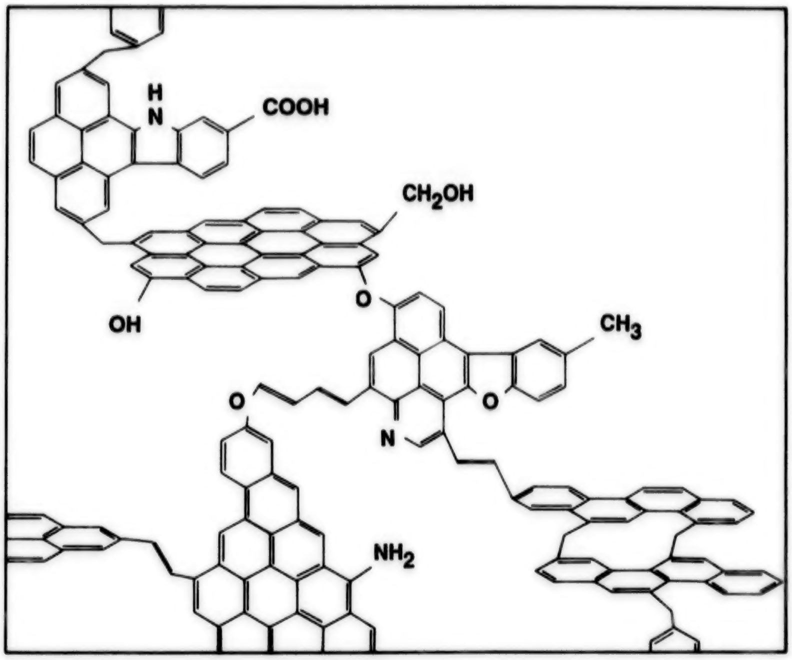


Figure 9-1. Schematic representation of the structure of kerogen-like material in carbonaceous chondrites. Note the three-dimensional array of aromatic moieties with aliphatic bridging units and side chains.

The salient properties of meteoritic organic matter include the following:

By far the dominant organic fraction is an insoluble macromolecular component, resembling terrestrial kerogen, that consists of polycyclic aromatic moieties distributed in a three-dimensional network with short aliphatic bridging units and side chains (fig. 9-1).

The soluble organic fraction contains many different classes of compounds, some of which, e.g., carboxylic and amino acids, reveal considerable, perhaps complete, structural diversity and a predominance of branched-chain molecules, whereas others, e.g., aliphatic hydrocarbons, do not.

The stable isotopes of C, H, and N are not distributed uniformly throughout the organic matter but show variations, some of which are probably related to the mode of formation of the organic molecules, whereas others probably reflect admixture of material from different sources. An example of the first kind is shown in figure 9-2, which shows the distribution of the C isotopes among the homologous series of alkanes and monocarboxylic acids: the systematic decrease in ^{13}C/^{12}C with increasing C number clearly reflects a synthesis in which higher homologs were made from lower by progressive

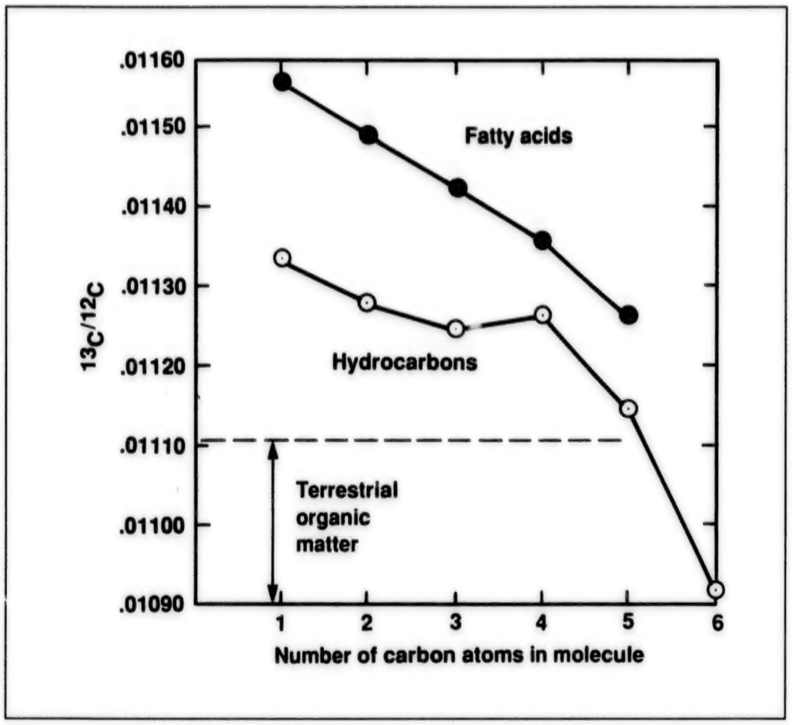


Figure 9-2. Distribution of carbon isotopes among the homologous series of alkanes and monocarboxylic acids in the organics of the Murchison meteorite. The systematic decrease in the isotope ratio with increasing C number reflects a synthesis in which higher homologs were made from lower by progressive addition of single C atoms.

addition of single C atoms. The effects of mixing discrete isotopic components are illustrated in figure 9-3, which shows C and H isotopic compositions of moieties identified within the kerogen-like material: the different components cannot be readily derived one from another by isotopic fractionation and are therefore believed to come from different sources.

Hydrogen in the kerogen-like fraction and in amino and carboxylic acids is considerably enriched in D relative to terrestrial H, and even more so relative to primordial solar-system H. These enrichments greatly exceed those that could be generated by known solar system processes and are therefore generally attributed to fractionation during ion-molecule reactions at very low temperatures in interstellar clouds (fig. 9-4).

The mode of synthesis is not the only interesting issue potentially addressable by the meteoritic record. The location and timing of that synthesis are also usefully, if not rigorously, constrained by the chondritic data. Thus, it now seems most likely that carbonaceous chondrites, like all other chondrites, are fragments broken off asteroids so that, given the apparently primordial zoning within the asteroid belt, a formation location for those meteorites inside the orbit of Jupiter is indicated. An origin for their organic matter in the inner solar system is thereby suggested, but the observed D enrichment mentioned above seems to require operation of a presolar mechanism, i.e., isotopic fractionation in interstellar molecular clouds. It is not known whether the interstellar H survived entry into the primordial solar system within the organic molecules or in some other form from which it was subsequently recycled into locally produced organic matter. For example, meteoritic amino acids are

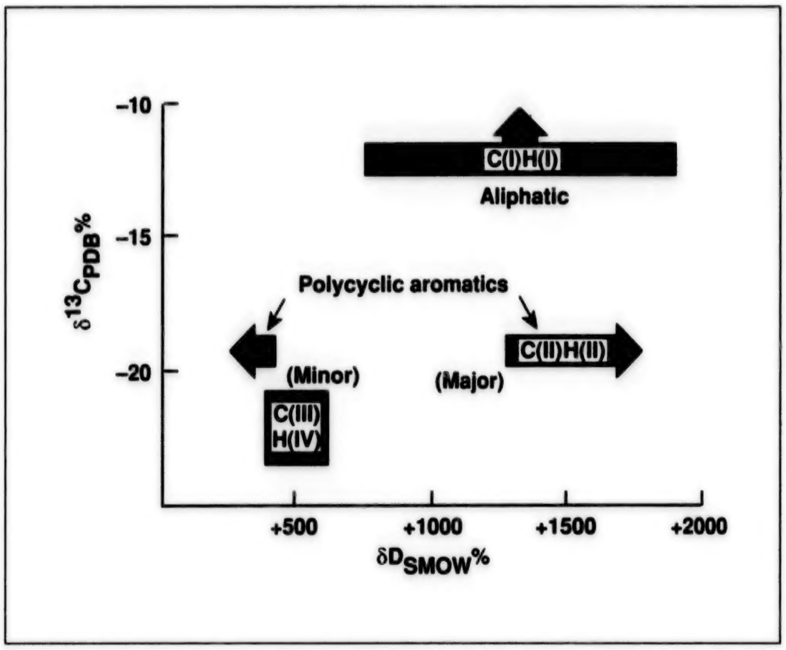


Figure 9-3. Isotopic composition of different moieties present within kerogen-like material in the Murchison carbonaceous meteorite. The values of the $^{13}C/^{12}C$ and D/H ratios are expressed in the delta notation as variations, in parts per thousand, relative to terrestrial standards: Pee Dee Belemnite in the case of C, Standard Mean Ocean Water in the case of H.

commonly believed to have been synthesized by the Strecker cyanohydrin reaction during aqueous activity in asteroid surface regions, and the precursor molecules for that synthesis have all been observed in interstellar clouds and, in the early solar system, could plausibly have had such an origin.

Radiometric dating places the epoch of aqueous alteration and impact-induced turnover on the parent asteroids within, at most, a few tens of Myr after solid formation in the solar nebula, and since it is most unlikely that the organic matter was introduced after that epoch, very early organic synthesis is indicated. This is obviously consistent with either interstellar, nebular, or planetesimal production of the organic matter.

Asteroids

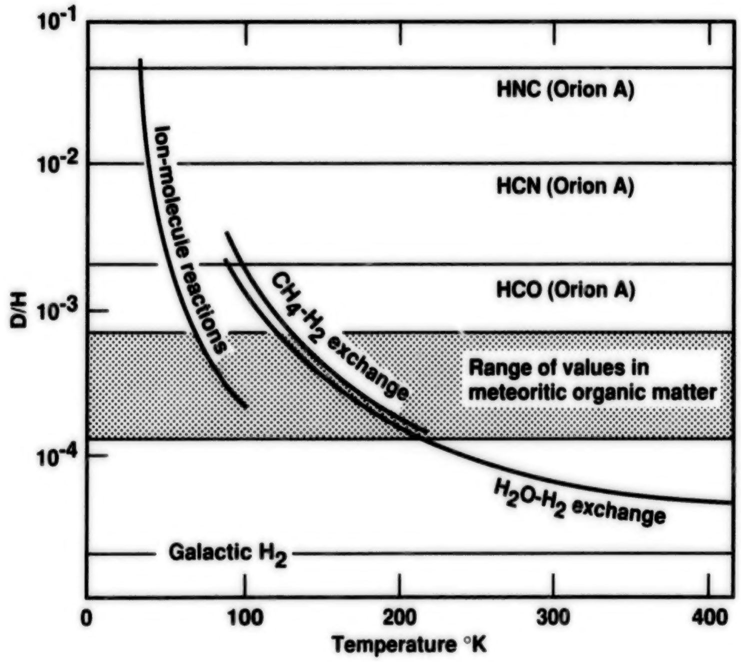
The albedos of the asteroids have been determined largely from infrared radiometric observations at 10 and 20 μm from Earth-based telescopes, and at 25 μm from the Infrared Astronomical Satellite (IRAS). On the basis of a modest sample of about 180 asteroids, ground-based radiometry established a bimodal distribution of asteroid geometric albedos, with peaks in the distribution at 0.04 and 0.15. The IRAS data base has yielded reliable albedos for over 1800 asteroids; in the sample of objects of diameter >40 km, both the bimodality and the median values of the two peaks are clearly confirmed, but in the sample with diameter <40 km, the distribution of albedos is more nearly Maxwellian with the peak at 0.05 (fig. 9-5).

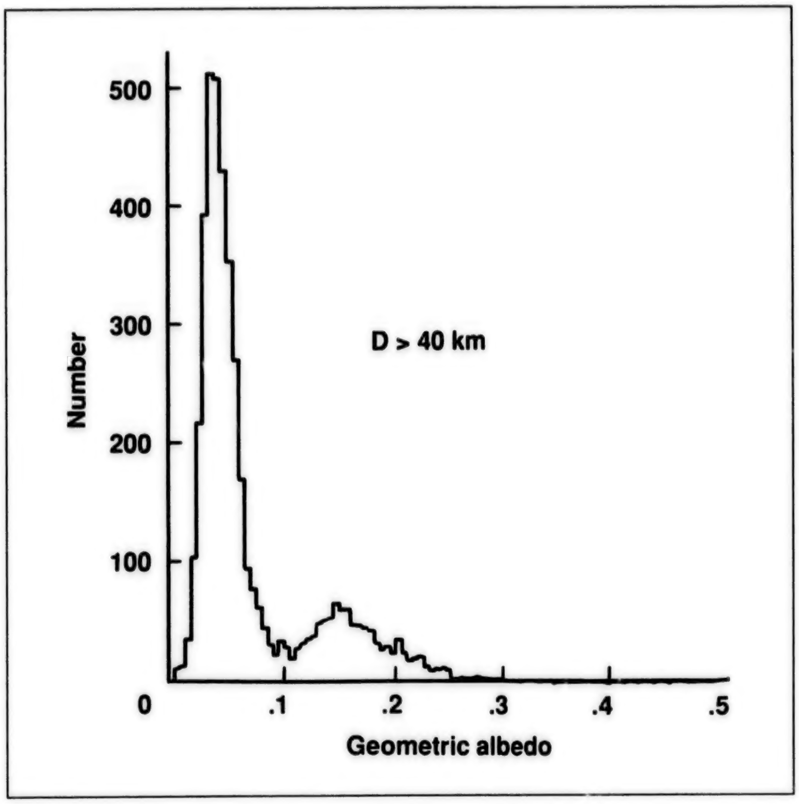
Telescopic observations have been used to classify the low-albedo asteroids into three basic categories according to their colors in the photovisual spectral region (0.3-1.05 µm). In this classification scheme, the C asteroids are usually neutral over most of the spectral region, with a downturn in reflectance toward the violet end of the spectrum. In the near-infrared (1-2.5 µm) the C-types are neutral to slightly reddish (a small upward slope in reflectance toward longer wavelengths). No discrete spectral absorption bands are seen in the C-type spectra in this region, but farther into the infrared some of the asteroids in this category have a distinct absorption band at 3 µm attributed to H_2O and OH in hydrated silicates in the regolith. Some of the C-types have no bound water, and there appears to be considerable variability among them, with several degrees of hydration represented. The presence of the 3-µm hydrated silicate band appears correlated with the strength of the violet downturn in reflectance, such that the "wet" C-types fit into a subclass called G.

The D-type asteroids are distinguished from the C-types by their significantly redder slopes both in the photovisual and near-infrared regions; the D-types include the reddest known asteroids. Hydrated silicates have not yet been found in asteroids of this class, but the D-types are difficult to observe because of their low albedos and their tendency to lie in the outer reaches of the asteroid system.

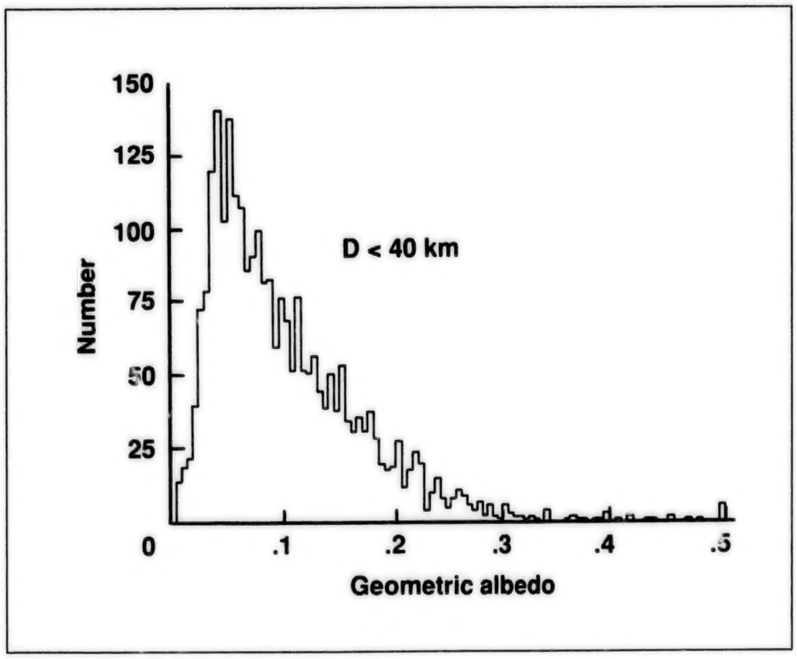
The P-type asteroids are similarly of low albedo, and may represent a transition between the C and D types. They have similar absorption in the UV and featureless red-sloping reflectance throughout the near-infrared. There is currently no information on the presence of hydrated silicates; as in the case of the Ds, they are very faint and difficult to observe.

Figure 9-4. Deuterium fractionation, relative to galactic H_2, caused by exchange reactions and ion-molecule reactions, compared with D/H values observed in meteoritic components and in the Orion interstellar cloud.





(a)



(b)

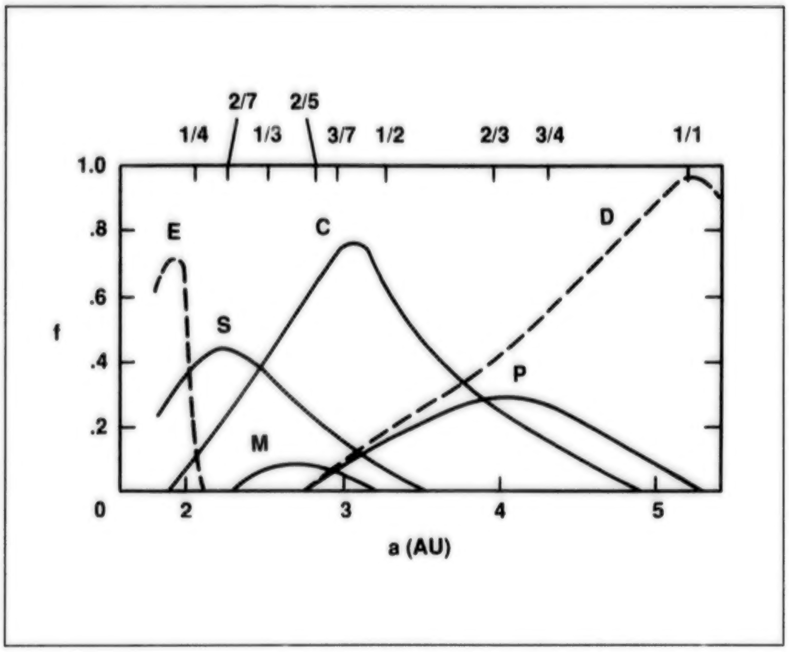
Figure 9-5. (a) Histogram showing the distribution of the number of asteroids of diameter >40 km with geometric albedo. These data were derived from measurements of the thermal radiation of the asteroids with the Infrared Astronomical Satellite (IRAS). The distribution is bimodal, with the greatest number at about 0.05. (b) The same plot as in (a), but for asteroids of diameter <40 km. The bimodality of the distribution is gone, but the peak still lies at about 0.05. This and figure 9-5a are adapted from E. F. Tedesco et al., "IRAS Observations of Asteroids," in Comets to Cosmology, Andrew Lawrence, ed., Springer-Verlag, 1988.

The distribution of the asteroid types with distance from the Sun has proved to be of some significance, once bias-corrected samples could be studied. A number of studies of the heliocentric distribution of various types that culminated in the appearance of a clear pattern are shown in figure 9-6. The middle of the asteroid main belt is at about 2.7 AU; inside this distance most of the asteroids are of the high-albedo classes, particularly the S-types (median albedo 0.15). Beyond 2.7 AU, there is a dramatic increase in the number of low-albedo objects; the C-types dominate the outer half of the main belt, and as they decline beyond 3 AU, the redder objects of the P and D classes increase toward their peak in the Hilda group at 4.0 AU. Color variations among the reddish types have been noted.

Two clear trends have emerged from this work. First, the low-albedo asteroids dominate the region beyond about 3 AU, and second, there is a strong trend toward increased redness of the spectral reflectances with increasing distance from the Sun.

Various investigators have suggested that the dominance of the low-albedo bodies and the increased redness in the outer portions of the asteroid system are results of a temperature gradient in the solar nebula that is preserved in the present distribution of the asteroids that formed at different distances from the Sun. The appearance of dark bodies at about 3 AU could represent the onset of formation in the solar nebula of carbon-rich minerals and macromolecular carbon compounds (organics) that color the materials of which the asteroids are composed. Alternatively, 3 AU could mark the inner limit in the solar nebula at which such organic compounds formed elsewhere could survive.

Figure 9-6. The distribution of a bias-corrected sample of asteroids by taxonomic type with distance from the Sun. This figure shows that the C-type objects peak in their distribution at about 3 AU, while the peaks for the P- and D-types are at larger distances. Types E, S, and M are higher-albedo objects not addressed in this paper. f is the estimated fraction of asteroids at a given distance in each class.



The possible presence of complex organics in the outer solar system bodies in the form of kerogen-like material, which compose the bulk of the organic matter in the carbonaceous chondrites (up to 70% in the CI meteorites such as Orgueil) was noted in 1980. Kerogen itself is a mixture of complex molecules which, in the case of terrestrial petroleum deposits, consists of a waxy substance that adheres to carbonate grains. At temperatures of about 750 K, the kerogen is altered to liquid petroleum. In the meteorites, it is intermixed with clay mineral grains and constitutes a very opaque, black component. Kerogen, or kerogen-like material, therefore, appears to be a plausible cause of the very low reflectivities of various solar-system objects, in part because few other natural substances are so black, and in part because the widespread occurrence of complex organic molecules is an acceptable working hypothesis, given its presence in interstellar space and the most primitive meteorites, and the relative ease of its genesis in plausible planetary environments.

One serious drawback to the kerogen hypothesis is that, by itself, kerogen, as prepared, for example, from coal tar has the wrong colors, although it is extremely low in albedo. For example, the photovisual reflectance of the very red D-type asteroids could be simulated by a synthetic mixture of clays and kerogens, the clays giving the red color and the kerogens the low albedo. The Trojan asteroid 624 Hektor was matched by a mixture of 85% montmorillonite clay and about 8-15% kerogen, plus some magnetite or carbon black. In part because the term kerogen carries a connotation of origin in terrestrial petroleum deposits, it is perhaps advantageous to substitute the term "macromolecular carbon" in the context of meteorite and related studies.

In laboratory reflectance spectra of carbonaceous chondrites the presence of bound water is revealed by a broad absorption band at 3 µm. On the long-wavelength portion of the featureless slope of this band, at 3.4 µm, there is superimposed a weak band complex assigned to the stretching mode in C-H in the organic compounds in the meteorite.

The 3.4-µm absorption band of C-H was sought in the reflectance spectra of asteroids that were selected on the basis of their bound-water band and their low albedo (i.e., the G-types). Tentative identification of the C-H band was made in asteroid 130 Elektra, based on the data shown in figure 9-7. In the laboratory spectra of carbonaceous meteorites, the C-H band is weak because of the high opacity of the finely granulated sample. This is true for pulverized samples of the bulk meteorite as well as for the concentrated insoluble organic residue remaining after removal of the other components. Similarly, the band is expected to be very weak in the diffuse reflectance spectra of an asteroid, making its detection quite difficult. The intrinsic weakness of the band, plus the inherent difficulties in registering the spectrum of a weak astronomical source in the 3-µm region, account for the uncertainties in the detection of C-H in 130 Elektra.

The importance of establishing the presence of organic molecules in the regoliths of asteroids is multifold. It is certainly connected with the general problem of the origin of specific kinds of meteorites, but it also bears on the chemistry of the solar nebula on the outer edge of the zone

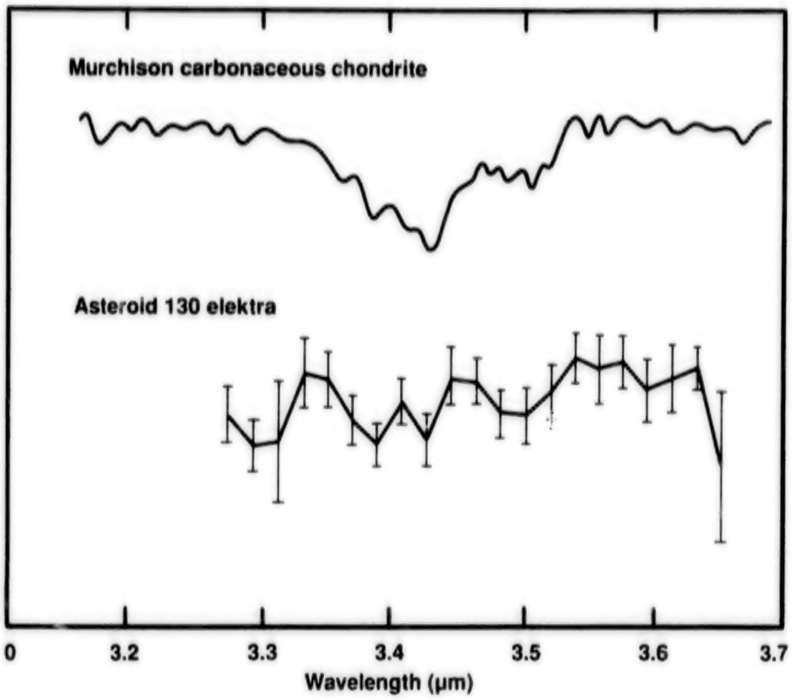


Figure 9-7. The reflectance spectrum of the G-type asteroid 130 Elektra in the region of the 3.4-µm C-H stretch band, with a laboratory spectrum of the Murchison carbonaceous meteorite for comparison. Effects of continuum slope due to sample and asteroid temperature as well as the strong band of bound water in each material have been removed. The correlation of the absorption features in the asteroid spectrum with the main feature in Murchison is taken as preliminary evidence for the presence of C-H on an asteroid. Adapted from D. P. Cruikshank and R. H. Brown, Science, vol. 238, 1987, p. 183.

of terrestrial-planet formation. In the broader sense, the question of the origin of macromolecular carbon molecules in the interstellar medium and incorporation into the solar nebula, versus the genesis of this material near the forming Sun is of great interest. The confirmed detection of organic molecular material on asteroids will be only the first step, to be followed by the identification of the species and the study of the space distribution of organic-bearing asteroids.

Low-Albedo Planetary Satellites

The odyssey of the Voyager spacecraft through the outer solar system has resulted in the discovery of a number of previously unknown planetary satellites, typically small and close to their parent planets. Together with low-albedo satellites known from telescopic discoveries, there are some 24 of these bodies of a few tens of kilometers in dimension. The impending encounter of Voyager 2 with the Neptune system carries the potential for the discovery of more of these small bodies.

The similarity in size, low albedo, and density (where known) of some of these satellites to dark asteroids has invited comparison and speculation on commonality of origin. The two satellites of Mars are often regarded as asteroidal in nature, though skepticism has been expressed on the issue of dynamical capture into their present circular orbits. The outer Jovian satellites are often compared with the Trojan asteroids (which share

Jupiter's orbit at the Lagrangian points). Dynamical considerations of the origin of planetary satellites show that asteroids or comets might be captured by planets under certain circumstances. Indeed, the only known way to acquire such retrograde satellites as Saturn's Phoebe and four of the outer satellites of Jupiter appears to be through dynamical capture. Thus, at least some of the small, dark planetary satellites may have originated in a way similar to the asteroids, though the details seem to be unique for each body.

In discussing the possibility of organic materials as the darkening component of the surfaces of asteroids, and particularly planetary satellites, it is worth bearing in mind the fairly low albedo of our own Moon. The overall geometric albedo of the Moon is 0.12, but that of the darkest lunar maria is about 0.07. Samples returned from the Moon attest to the fact that the regolith on the lunar maria has no organic component, and the low albedo is attributed to the dark minerals and texture of the regolith. It is very difficult to make natural surfaces with an albedo half that of the lunar maria without resorting to some other component, and macromolecular carbon compounds are a plausible choice.

The physical evidence for a possible macromolecular carbon component of planetary satellites consists of broad-band photometric measurements of the colors of these objects (0.3-3.5 μm), and in some cases actual spectra over limited wavelength intervals. The limiting factor in obtaining spectral data is the faintness of small planetary satellites, even when observed with the largest telescopes and most sensitive spectrometers. Spacecraft sent to the planets with small dark satellites have not carried spectrometers suitable for studies of their small dark satellites, with the result that only color data (from space and from the Earth) and imperfect spectra (from the Earth) exist. An added inconvenience is the absence of strong diagnostic spectral features in accessible spectral regions, owing partly to the nature of organic molecules themselves and to the intrinsic weakness of most spectral features of low-albedo solids seen in diffuse reflectance. The most promising spectral band is that of CH at 3.4 μm seen in some carbonaceous chondrites, and possibly in asteroid 130 Elektra, as noted above. This band has not been observed on any planetary satellite (except Titan), though attempts have been made to detect it on Iapetus, as we discuss below.

The striking bimodal distribution of a very low-albedo surface unit and an expanse of water ice or frost on Saturn's satellite Iapetus has commanded special attention in the search for organic material on bodies in the outer solar system. Data from Voyagers 1 and 2 confirmed the pattern of distribution of the dark material that had been derived from telescopic studies showing that the dark covering on the leading hemisphere of the satellite is centered on the apex of the orbital motion. Such a geometry argues strongly for a method of emplacement of the dark material that depends in some way on the influx of material from space, though the details are unclear.

There are two main ideas concerning the origin and composition of the dark material on Iapetus. In one, dark material is emplaced by deposition from a source outside the orbit of Iapetus, or it results from a process of impact concentration to the surface of dark material already present and distributed in the upper layers of the satellite. In another model, black macromolecular carbon is formed *in situ* from methane clathrate by solar UV bombardment. The former model appears to be favored in view of the telescopic data and the spacecraft images showing the manner in which the surface deposits relate to topography on Iapetus.

The dark hemisphere of Iapetus, which is seen from the Earth face-on for an interval of about 48 hours once during each 80-day orbit of the satellite around Saturn, is sufficiently bright to permit low-resolution spectrophotometric observations in the photovisual and near-infrared spectral regions. Such data have been compared with laboratory samples of candidate materials, with concentration on materials of very low albedo and red color.

It is important to remember that very small quantities of dark, opaque powder can have a strong effect on the albedo of the material in which it is incorporated. For the case of the dark hemisphere of Iapetus, for example, it has been demonstrated that the abundance of water ice on that hemisphere could range from less than a few weight percent to as much as 95 weight percent. Finely divided dark material, such as charcoal, lampblack, magnetite, or macromolecular carbon, is extremely effective in reducing the overall albedo of the mixture and in quenching the strong ice absorption bands normally seen in the near infrared.

When fully corrected for the effects of the minor contamination of the dark hemisphere with relatively pure water frost in the polar regions, the reflectance of Iapetus between 0.3 and 2.5 μm is relatively smooth and strongly upward sloping toward longer wavelengths (that is, it is red), as seen in figure 9-8. Comparisons were made using this spectrum of Iapetus and laboratory data for the organic component of the Murchison CM carbonaceous chondrite, and an insoluble coal-tar kerogen. Neither the Murchison organic extract nor the coal-tar kerogen had the correct spectral reflectance, but mixtures of the kerogen with clay were produced and a good match in both spectral reflectance and albedo was achieved. An organic-clay mixture with about 10 weight percent kerogen produced the best match, as shown in figure 9-8.

The 10 newly discovered small satellites of Uranus all orbit very close to the planet. Only one, Puck, was imaged at sufficient resolution by Voyager 2 to show surface relief and its slightly nonspherical shape. The albedos are all very low, in the range of 0.04-0.06, and comparison is often made with the particles in the Uranian rings, which are also quite black. Furthermore, the ices of the five major satellites all appear to be contaminated to some degree with neutral material of low albedo which may have the same origin. Whether or not this material has any relationship to macromolecular carbon

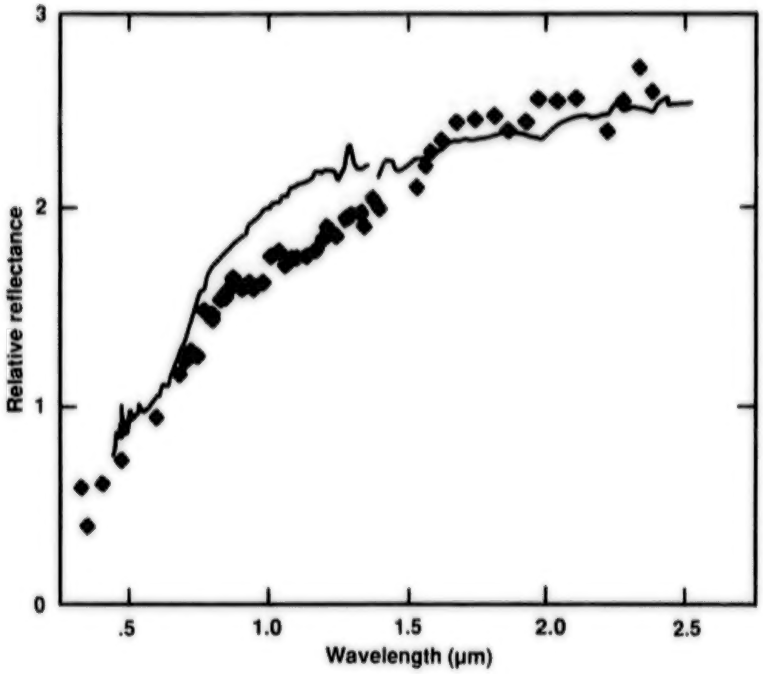


Figure 9-8. Spectral reflectance of the low-albedo hemisphere of Saturn's satellite Iapetus (data points) in comparison with the reflectance of a laboratory sample of 90% clay plus 10% organics (line). The general shapes and absolute values of the spectra are very similar over this broad wavelength range. From J. F. Bell et al., Icarus, vol. 61, 1985, p. 192.

cannot be established at this time, but it has been shown that irradiation of surface ices by the ion and electron fluxes observed by Voyager at Uranus can darken the surfaces if those ices contain some carbon. While the presence of water ice on the larger Uranian satellites is well established, the telescopic search for other substances, such as CH_4, NH_3, CO, and CO_2 has so far only given a faint hint of NH_3.

Pluto, Charon, and Triton

The low temperatures at the surfaces of the outermost planets and satellites permit methane and some other simple hydrocarbons to exist as solids, though the vapor pressures are sufficiently high that a certain mobility (sublimation and recondensation) is expected in seasonal cycles. Methane has been found on Pluto and Triton, and there is strong evidence for condensed nitrogen on Triton. Water ice has been found on Pluto's satellite Charon, which has led to some early models of the escape of original methane from the satellite.

While neither solid methane nor condensed nitrogen (solid or liquid) are quite neutral in the photovisual region of the spectrum (0.3-1.0 μm), both Pluto and Triton are notable for their yellow-red colors. Pluto's surface is clearly nonuniform in albedo and color distribution, with a concentration of relatively darker material at the equator and high-albedo deposits at the poles. There is some suggestion of surface variegation on Triton, but the issue is confused by an apparent change in both the spectrum and the photometric lightcurve that occurred sometime in the early 1980s.

Laboratory Studies

Numerous laboratory studies in the last few years are pertinent to the outstanding questions of the nature and origin of the very dark material on surfaces of some solar system bodies. The experimental work falls into two general categories: (a) spectroscopic and photometric studies of natural materials in various mixtures, including extracts from meteorites, and (b) similar observations of synthetic materials produced or altered through the effects of irradiation by ions, protons, electrons, ultraviolet, or gamma rays.

Radiation-processed ices and organic materials have been studied in several laboratories. Organic residues from carbon-containing gases or ices have been produced by electrical discharges, UV radiation, high-energy ions, and fast protons. Some preliminary planet-related experiments with gamma-irradiated mixtures of nitrogen and methane have also been done in the laboratory. In addition, direct studies of meteoritic organics in the context of the interstellar and planetary occurrence of this material have moved steadily forward.

Ices of planetary significance (H_2O, CO_2, CO, CH_4, NH_3, etc.) all show diagnostic spectral features in diffuse reflectance, and a number of these solids have been identified on cold planetary bodies. Similarly, hydrous clays and igneous rock-forming minerals of planetary significance (pyroxene, plagioclase, orthoclase, and olvine) are identified on planets, asteroids and satellites through their spectral absorption features. The laboratory work has stressed spectral regions where actual astronomical data exist or can be obtained within the constraints of flux level and transparency of the Earth's atmosphere. Extension of the laboratory data into spectral regions from which spacecraft data will eventually come has only begun.

New work is in progress on combinations of ices and minerals, which must occur because the overall albedos of some icy bodies are too low and the colors are incorrect for pure ices. Laboratory work is being undertaken in support of the computation of synthetic spectra of pure minerals and ices and, most significantly, combinations of these materials in a range of grain sizes and surface textures, through knowledge of the theory of scattering from planetary surfaces and detailed knowledge of the optical properties of the candidate materials. Eventually, this computational technique can be extended to mixtures of organic solids with ices of planetary interest.

Work on organic compounds with application to planetary problems is being undertaken in the context of cometary nuclei and solids in the interstellar medium. Coloration and darkening of methane clathrate and other ices has been produced by charged-particle irradiation. Yellow and brown organic products are generated when otherwise white ices of various combinations of water, methane, and ammonia are irradiated for times that are relatively short in terms of equivalent exposures in space. Similar experiments have been conducted with the production of non-volatile colored organic material from irradiation of methane ice. UV irradiation of ices containing methane and other simple hydrocarbons can produce more complex molecules, in particular a variety of polycyclic aromatic hydrocarbons, which show spectral features giving a close match to infrared bands in the interstellar medium and in some interplanetary dust particles.

In some cases these experiments produce very low-albedo material that is a plausible constituent of the dark planetary surfaces and meteorites discussed in this review, and in other cases, colored material of possible relevance to the yellow surfaces of Pluto and Triton are formed. The relative ease with which complex organics are produced from the simple raw materials found in many places in the outer solar system virtually ensures a significant role for them in the surface chemistry of the comets, asteroids, planetary satellites, and some planets. The current production of stable and complex hydrocarbons in the atmospheres of the outer planets has been appreciated for quite some time.

Summary

Low-albedo bodies in the solar system appear to come in a range of colors, with Iapetus being very red and the outer satellites of Jupiter being nearly neutral (nearly flat reflectance curves). This statement also applies to the asteroids in the outer parts of the main belt and in the Trojan groupings, and it is now accepted that the surfaces of many comet nuclei are extremely low in albedo. Macromolecular carbon, which is found in meteorites and interplanetary dust particles, is a plausible low-albedo constituent of these dark asteroid, comet, and satellite surfaces; indeed, the presence of organic matter in cometary dust has been established.

Insofar as the asteroids are the source bodies of the carbonaceous meteorites, the presence of organic material on at least some classes of asteroids is a certainty. Telescopic searches for the CH spectral signature strongly suggest its presence on dark, hydrous asteroids, but confirmation and further exploration are needed. More information is needed on the relationship between carbonaceous chondrites and specific asteroids or classes of asteroids. Major questions remain about the origin of the meteoritic organic matter, particularly concerning the role played by presolar, interstellar processes. This issue will be greatly clarified by detailed analysis of the organic population of comet nuclei, in which interstellar molecules should be preserved. In that connection, the Comet Rendezvous Asteroid Flyby (CRAF) and Comet Nucleus Sample Return Missions will be of prime importance.

Solid organic matter is presently being synthesized in the atmospheres of the methane-bearing planets and satellites of the outer solar system, and probably on the solid surfaces of Pluto and Triton. Photochemical smog is continuously produced in the atmospheres of Titan and the giant planets, and the presence of methane frost or ice on the yellow-colored surfaces of Pluto and Triton virtually ensure the alteration to more complex organic solids by the action of UV and the galactic cosmic ray flux.

More generally, we wish to know which chemical processes have been effective at synthesizing organic molecules from simple inorganic gases, and what factors have controlled the extent of chemical, pre-biotic evolution on different solar system bodies. For example, chemical evolution led to the emergence of life on Earth but not, apparently, on the carbonaceous chondrite parent asteroids. Were some components of the "primordial soup," present on Earth, missing on the asteroids, and if so, which?

Additional Reading

The text and references for this article were up to date at the time of writing in 1988. Many additional papers on related topics have been published since that time; references to two of them are cited here.

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Comets: Role and Importance to Exobiology

N93-18556

A. H. Delsemme

Comets have been repeatedly linked with exobiology for the following reasons:

1. Among the heavenly bodies, they contain the largest amount of organic molecules;

2. The observed comets are all in unstable orbits; since they cross the planetary orbits, orbital precession guarantees that sooner or later comets hit the planets. Hence they are a possible source of water, carbon, and organic molecules for the planets;

3. Comets decay into gas and dust; this is a source of interplanetary particles that may also reach the planets. As expected, interplanetary particles collected high in the atmosphere of the Earth also contain organic molecules;

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4. Comets show a strong analogy with carbonaceous chondrites; as a matter of fact, if a large carbonaceous chondrite were on a cometary orbit, it would be indistinguishable from a comet. In particular, it would develop a cometary tail, and the spectrum of its head would also be indistinguishable from a cometary spectrum;

5. The previous analogy has been extended to Mg 26 (extinct Al 26) observed in many chrondrules. If also present in comets, Al 26 would have produced a transient source of heat (it decays with a half-life of 0.7 million years) large enough to keep water in the liquid state in the cores of comets for a time long enough to make complex organic syntheses from prebiotic molecules possible;

6. Fossil evidence confirms a chain of events suggesting that comets are the major source of the oceans and of the atmosphere of the Earth, as well as that of the organic compounds that were needed to get life started on the Earth;

7. The previous chain of events can be traced backward to interstellar space, namely to those dense interstellar clouds that contain numerous organic molecules produced by ion-molecular reactions. Comets seem therefore to be the cosmic messengers that have transported organic molecules from the interstellar space down to Earth, mostly during the late stages of the accretion of the terrestrial planets.

The understanding of the chemical nature of the cometary nucleus has considerably progressed recently, in particular thanks to the multiple flybys of comet Halley. Most of the questions of exobiological interest have therefore received much attention, but none have been unequivocally answered. The basic reason is that the relevant molecules were damaged by the high velocity of the flyby prior to analysis. What we know now about cometary chemistry is reviewed in the following section.

In the last decade, considerable progress has also been achieved in the understanding of the chain of events whose outcome was the biosphere of the Earth. Crater statistics on the Moon have not only numerically confirmed the early intensity of the cometary bombardment of the Earth, but numerous other fossil data from meteorites (mentioned later) are consistent with the sequence of earlier events that explain the mechanism of this bombardment. It will be shown in particular how the general scenario describing the origin of the solar system also explains the exogenous origin of the terrestrial planets' atmospheres.

Since organic molecules are ubiquitous in the Cosmos, cometary processes that bring organic molecules from space to rocky planets seem to be an integral part of the chain of events forming planetary systems, a conclusion of far-reaching importance for exobiology.

Chemistry of the Cometary Nucleus

Radial Homogeneity of Nucleus

Quantitative knowledge about cometary chemistry has mostly been derived from the bright comets of the 1970s and complemented by the wealth of information brought about by the flybys of comet Halley during its 1986 passage. However, other historical comets have displayed enough diversity to suggest the existence of large physical and chemical differences.

The author reviewed this matter in 1987 and has come to the conclusion that in spite of their apparent diversity, comets represent a population of essentially homogeneous objects. In 1982, he also listed circumstantial arguments that all favor undifferentiated nuclei. The gist of the arguments is that, if nuclei were radially differentiated, their decay with aging or their fragmentation by splitting would sooner or later produce observable changes. However, all observations of cometary material that originated at different depths in the pristine nucleus show the same properties, namely:

1. The same dust-to-gas distribution pattern;

2. The same spectral composition of the volatiles;

3. The same structural strength against fragmentation;

4. The same vaporization pattern after fragmentation.

Of course, minor heterogeneities would not be observable below a certain size (30 to 50 meters). There are, however, surface differentiation effects. They probably preexist in "new" comets, because of their surface irradiation by galactic cosmic rays. Of course, as soon as a comet penetrates the inner solar system, the outgassing of the surface layers and the sintering of dust into a crustal slag enhances the surface differentiation considerably.

The apparent diversity of the cometary population comes more from the nucleus fragmentation and decay, than from any physical or chemical difference in the pristine component.

The Dust-to-Ice Ratio

By definition, a "primitive" mixture is the condensate at 30 K of a gas of solar composition. At such a low temperature, hydrogen, helium, and neon (that is, the most volatile gases) do not condense. Therefore, they are considerably depleted in a primitive mixture, whereas all the other elements have remained in solar proportions; in particular, carbon, nitrogen, oxygen, and sulfur are not depleted. It is worth mentioning that the most primitive carbonaceous chondrites do not come straight from a primitive mixture, since their carbon, nitrogen, oxygen, and sulfur are depleted in variable amounts. In a primitive mixture, all metals including silicon, iron and magnesium are assumed to be more or less oxidized into silicates. The amount of oxygen used for that purpose depends somewhat on the nature of the silicates: the fraction oxygen/metals may go from a low 4/3 for olivine (in numbers of atoms) to a high 9/5 for serpentine. Assuming that the rest of the oxygen as well as C, N, and S is in the volatile ices, then the primitive dust-to-ice mass ratio goes from 0.54 to 0.66 if Cameron's (1973) abundance table is used; it goes from 0.61 to 0.76 if we use strictly solar data. Including possible uncertainties, the primitive ratio remains in the range of 0.65 ± 0.11, if *all* the carbon is in volatile compounds (like CH_4, CO, or CO_2). If 3/4 of the carbon is in refractory compounds, as was observed in the dust grains of comet Halley, then the "primitive" ratio becomes 1.10 ± 0.22. To summarize the situation, depending on where most organic molecules go (in the refractory dust or in the volatile ices), the primitive dust-to-ice ratio may vary from 0.5 to 1.3.

The notion of primitive ratio is useful because it sets a lower limit for the dust-to-ice mass ratio. If comets condensed or accreted at a very low temperature from a reservoir of solar abundances and were never processed by heat later on, then the dust-to-ice ratio is the primitive ratio. If comets were heated during the course of their evolution, they lost more gas than dust; some of the gas may also get polymerized into dust; hence, the dust-to-ice ratio becomes larger than the primitive ratio. This is likely to occur locally, in the outer layers of the nucleus, first after their irradiation with cosmic rays, second in the crustal slag after perihelion passage. It is difficult to conceive any phenomenon that could diminish the primitive ratio by losing more refractory material than volatile stuff.

The actual dust-to-ice ratio of pristine comets is difficult to establish, because the surface decay of the nucleus alters the ratio. However, the interpretation of large outbursts suggests that the sudden blow-off of a mantle of dust often occurs. Ratio fluctuations out of steady state have been reconstructed day after day from data obtained during the outburst of comet Arend-Roland. After the outburst, the dust-to-gas mass ratio seemed to go back asymptotically to its steady-state value; for this reason it is believed that the steady state is representative of the inner composition of the nucleus. The best values for the steady-state dust-to-gas ratio have probably been obtained by the Finson-Probstein method for two comets only. They are 0.8 ± 0.2 and 0.6 ± 0.4 for comets Arend-Roland and Bennett, respectively. Because of the error bars, the results are not essentially different and are consistent with a primitive ratio in each case.

The Vega experiments have found 5 to 10 tons/second of dust and 30 tons of gas for the production rate of comet Halley, but a study of the tail isophotes has shown that the dust-to-gas ratio during the flybys was far from the steady state. At any rate, the discrete data of the Dust Impact Detection System (DIDSY) carried by the Giotto Spacecraft suggest the existence of a large population of massive particles not taken into account in the production rate of dust found by the Vega experiment. Comet Halley's dust-to-ice ratio remains therefore remarkably consistent with a primitive ratio.

The dust-to-gas mass ratio in the coma of 17 comets has been studied, and very low values, between 0.05 and 0.30, were found. Of course a comet, even with a primitive ratio, may vaporize more gas than dust when not at a steady state, but this cannot last for a long time without choking the vaporization under a growing mantle of outgassed dust. On the other hand, it is unlikely that the 17 comets all showed lower values than the steady state. Assumptions were made: an albedo of 0.05 for dust and a nominal expansion rate of

C-3

Table 10-1: Dust in Comet Halley.
Mean Chemical Composition in Mass %

Organic 33%		Inorganic 67%	
Unsat. hydroc.	16.0%	Silicates	51.5%
H, C + O	5.2%	FeS (troilite)	6.0%
H, C + N	4.5%	C (graphite)	3.0%
H, C + S	1.8%	S (sulfur)	1.0%
Water	5.5%	Water	5.0%

1 km/second for gas. Additionally, an assumption was made on the grain size distribution of the dust that was not measured, since the Finson-Probstein method was not used on the dust tail isophotes. Since the existence of a steady state was not established for any of the cases, no hard inconsistency has been found with a primitive dust-to-ice ratio of 0.5 to 1.3 for the 17 comets in question.

Nature of the Dust Grains

Before the flybys of comet Halley, Brownlee particles collected by NASA's U2 plane in the upper atmosphere were probably the most reliable source of information on the cometary grains. Their analysis suggested that they were similar to carbonaceous chondrites, containing, however, more carbon and sulfur. Typically, the particles contained 5% C by mass, instead of 2.5 and 3.2%, the median, respectively, for CM and CV chondrites.

Yet the balance sheet for carbon in comets suggested that 3/4 of the cosmic carbon was still missing. At that time, I proposed that cometary dust could contain up to 30% carbon by mass, in order to reach cosmic abundances. The dust results of the Vega 1 flyby of comet Halley have supported this view rather well, although they revealed that the individual dust grains vary greatly in mass and in composition. A chondritic silicate fraction without carbon seems to form the grain core, which is embedded in a variable amount of organic material. Most of the grains are from 0.2 to 1.0 µm in diameter, with a steep decrease in the number of larger-sized grains. Table 10-1 represents the mean composition averaged over 80 particles.

The bulk of the carbon is, of course, in the organic fraction; an extra amount of 3% reduced carbon, presumably graphite, is also counted with the inorganic fraction. With this amount of carbon in the dust, the problem of the apparent carbon deficiency in comets seems to have completely vanished.

The inorganic fraction (67% in mass) contains 51.5% silicates (mainly of magnesium and iron), 6% troilite FeS, 3% carbon, 1% sulfur, and 5.5% water.

The organic fraction contains 16% hydrocarbons with an H/C ratio close to 2/3, suggesting unsaturated hydrocarbons like C_3H_2; there are also 5.2% of oxygen-containing species, probably formic acid and formaldehyde in about equal amounts, and 4.5% nitrogen-containing species, a mixture of hydrogen cyanide, acetonitrile, aminoethylene, pyrrole, pyridine, pyrimidine (50 coincidence pairs in mass spectra); purines and adenine would also be present in several grains, whereas there is no indication for amino acids; finally, there is 1.8% of sulfur-containing species. Some molecules contain several oligo elements; amino methanol seems to be present, as well as some O's that seem to be occasionally exchanged by an S atom in several organic molecules.

If recently published identifications are all correct, they imply very important conclusions for exobiology, namely:

1. The prebiotic molecules HCN, H_2CO, and H_2O are present;

2. The precursors of the nucleic bases (purines and pyrimidines) are probably present;

3. The amino acids are apparently absent.

However, insufficient details are available to check the validity of the method used. The contentious nature of the results (in particular items 2 and 3 above) stems from the 78 km/second velocity of the flyby. The high-velocity impact of the dust grains on the neutral mass spectrometer not only vaporizes grains but destroys molecules into atomic and molecular fragments. It is surprising that during impact large condensed heterocycles retain a recognizable spectral pattern of molecular fragments, but this has been verified in the laboratory for velocities up to 10 km/second. At any rate, the matter is pending if only because of the dramatic significance of the possible results.

Thanks to the large amount of organic material in the dust of comet Halley, there is no doubt now that the abundance of all elements, including the light elements C, N, O, and S, are present in almost cosmic proportions in this comet, with the exception of H (and presumably undetected He and Ne), depleted by a factor of about 500. The same situation is likely to exist in other comets.

Another important conclusion is the presence of a large fraction of rather volatile molecules in the CHON particles: 15 to 30% of the so-called "dust" is probably formed with "icy" molecules that may be almost as volatile as, and even possibly more volatile than, water; therefore they must vaporize according to the "icy grain halo" model. In particular, their decay may explain the existence of an extended source of some of the gaseous species observed around the nucleus, and extending into the coma.

For example, formaldehyde has been observed in the CHON particles (those particles that contain only C, H, O, and N atoms); its mass seems to represent several percent in mass of the total dust. Formaldehyde is more volatile than water; it has also been identified in the gaseous coma; its photodissociation by the solar light takes place in a few 10^3 seconds; and its branching ratio into CO plus other fragments (H_2 or H + H) is 69%. However, it can be freed by grains with a lifetime of a few 10^4 seconds and it can explain the observed extended source of CO.

Table 10-2: Light Elements in Some Bright Comets. Mean Abundances in Volatile Fraction

Element	Bright comets	Comet Halley
H	1.8 ± 0.4	1.9 ± 0.4
C	0.20 ± 0.10	0.20 ± 0.05
N	0.10 ± 0.05	0.10 ± 0.05
O	1.00	1.00
S	0.003 ± 0.002	0.01 ± 0.005

All elements are normalized to oxygen = 1.0.

Origin of the Dust in Comets

Some cometary grains contain metals or silicates only; some other grains contain a silicate core plus a variable mantle of organic molecules. Finally, the CHON grains do not contain metals at all, only organic molecules. The grains—most in the 0.2 to 1.0 μm range—are sufficiently different chemically to imply that they were condensed at different places and brought together (before or during accretion) by a process comparable by all means to sedimentation. This is rather similar to the carbonaceous chondrites which also contain a mixture of fine grains of different origins, namely high-temperature and low-temperature grains, oxidized and reduced grains, hydrated and anhydrous grains, etc.

The most plausible place for this sedimentation to happen has been described in models of the solar nebula. When the turbulence subsides because mass transfer to the accretion disk has stopped, then dust separates from the gas and sediments into equatorial rings, thereby preparing the stage for the accretion of planetesimals. In this scenario, the dust of carbonaceous chondrites has sedimented in the outer asteroid ring and the frosty and icy cometary dust has sedimented at distances between those of Jupiter and Neptune or beyond.

The recent observations of the chemical diversity of the dust in comet Halley seems to be consistent with such a scenario. Recent proposals that reconsider alternate origins for comets may have difficulties in explaining the chemical diversity of their dust grains.

Nature of the Volatile Fraction

There is a deep similarity in the spectra of many comets in the visual and the beginning of the ultraviolet. Many early differences came from observational artifacts (spectral and spatial dispersion, sensitivity limits of the photographic plate, etc.). The same remark has been expressed for the vacuum ultraviolet, suggesting that there might be a deep similarity in the composition of most comets.

It has been established that comets Bennett 1970 II, Kohoutek 1973 XII, and West 1976 VI were similar in many respects. After careful discussion, the elemental ratios of H, C, N, O, and S in their volatile fraction were found to be close enough to be incorporated into a single model, representing "the average bright comet of the 1970s." The final model is given in table 10-2.

It is worth mentioning that the elemental ratios of comet Kohoutek were oscillating wildly; the observations had picked up epochs close to two outbursts and far from steady state. It was, however, possible to make use of the observations. An outburst propagates in a few days into the coma; because of their longer photodecay lifetimes, the H atoms had been produced an average of 11 days earlier than the OH radicals observed simultaneously. Correcting for the delays damped the spurious fluctuations due to the outbursts and yielded a ratio H/OH = 1.8 ± 0.8 for the same production epoch of H and OH. In the same way, when C and O were corrected for the two different ionization lags of 2 days and 5 days respectively, the C/O ratio yielded an average of 0.22 ± 0.06 for Kohoutek (close to 0.21 ± 0.03 for the comet West), making the two spurious ratios of 0.08 (January 5) and 0.58 (January 8) disappear. This example has been discussed in detail to show that early differences came more from deficiencies in the models than from actual compositional differences.

The complete quantitative molecular analysis of the volatile fraction remains, however, a difficult feat, because the number of species involved remains uncertain, and the excitation mechanisms that bring different molecules into visibility in the spectra remain intricate. As far as the traditional bands observed in the visual are concerned, spectral differences from comet to comet are small and concentrate around the same "normal" ratios, for instance, $C_3/CN = 0.2$ and $C_2/CN = 1.5$.

The situation is somewhat different in the vacuum ultraviolet, where major carbon compounds seem to be variable in some comets. The diversity of the CO/OH ratio in comets Kohoutek 1973 XII, West 1976 VI, and Bradfield 1979 X has been mentioned in others' work. The interpretation of the two outbursts of comet Kohoutek given in the previous paragraph is convincing enough not to attach any significance to the spurious value 0.58, but the very low value of CO/OH = 0.044 of comet Bradfield has not been explained. It could come either from the same type of fluctuation out of steady state, or from genuine elemental differences.

It has been suggested that the variability of the CO/H_2 ratio is large: 27% for West, 18% for Halley, and 1% for Bradfield. The differences do not necessarily reflect intrinsic compositional differences between comets, but may come from deficiencies in the models used to interpret the observations.

The CO Distribution in Comet Halley

Many observations have been collected on or near the dates of the comet Halley flybys. For this reason, the CO distribution in comet Halley needs a separate discussion. Not surprisingly, important discoveries come from the flybys themselves.

The Giotto neutral gas mass spectrometer has discovered the existence in the coma of a large source of CO extending up to 15,000 km from the cometary nucleus. Independently, a large organic fraction containing, in particular, formaldehyde (H_2CO) and formic acid (HCO-OH) was discovered in the dust grains analyzed by the dust impact gas spectrometer PUMA of the Vega 1 spacecraft. Although the two groups have not yet mentioned the correlation between these two discoveries, we will verify in the next section that the photodissociation of H_2CO and of HCO-OH from the volatile fraction of the grains explains quantitatively the origin of the extended source of CO. Other explanations do not seem therefore to be justified.

From Giotto data reported in 1986, the punctual CO source represents 40% and the extended source 60% of the total CO flux. A total production of $CO/H_2O = (10 \pm 5)\%$ is given. The error bar is wide because other molecular fragments might contribute to the 28 a.m.u. peak. Since N_2 is the other major contributor to the peak, we have used the feedback of the very low evaluation of the total elemental N coming from other data (table 10-2), to deduce that the actual value of CO/H_2O lies in the upper range of limits, and can be reasonably represented by $CO/H_2O = (13 \pm 3)\%$.

The other source of information on CO comes from the vacuum ultraviolet spectra, from which $CO/OH = (10 \pm 5)\%$ was reported; however, because of the discovery of formic acid whose photodissociation represents another source of OH, we must remain aware that more than ever OH is not an automatic measure of H_2O. The CO = 10% reported from the Pioneer Venus spacecraft should not be construed as an identical result to the vacuum ultraviolet spectra results. The carbon resonance line used for the previous ratio represents only somewhat more than one-half of the total carbon, because the rest has remained in ionized CO^+ blown away at a high velocity by the solar wind.

Finally, other researchers give $CO/H_2O = (20 \pm 5)\%$ on February 26, 1986, and $(17 \pm 4)\%$ on March 13. Only the latter is close enough to the Giotto flyby to be compared and used here. We conclude this discussion by adopting

$$CO/H_2O = (13 \pm 3)\%$$

as the only possible value of the ratio which is reasonably consistent with all observations, including vacuum ultraviolet as well as mass spectra.

Dust Contribution to the Volatile Fraction

The large organic fraction of 33% (in mass) of the dust, found in the grains of comet Halley, brings new problems and new vistas related to the interface between dust and volatiles.

First, the organic fraction cannot be completely refractory, otherwise it would have survived the entry into the upper atmosphere and would therefore still be present in the pores of the Brownlee particles. The interpretation of recent data yields a total carbon content of 25% by weight, versus 5% in particles. The missing 20% carbon by weight cannot be too volatile either, otherwise it would have vaporized into the coma before being detected by the Giotto's neutral mass spectrometer. Table 10-3 suggests that it is combined with about 10% by weight of H, O, N, and that the fraction represents a total of 30% of the dust; in particular, 5.2% (by weight) of the dust seems to come from a couple of major constituents: formic acid and formaldehyde.

Table 10-3: Elemental Abundances in Comet Halley

Name of element	Dust (mass)	Gas (mass)	Total (mass %)
H	7.8	21.9	5.3%
C	72.0	29.1	18.1%
N	2.8	16.9	3.6%
O	104.0	194.1	53.4%
S	16.0	3.8	3.5%
Mg	17.0	—	3.0%
Si	28.0	—	5.0%
Fe	39.1	—	7.0%
A + Na	1.7	—	0.3%
Ca, Cr, Mn, Ni	4.0	—	0.7%
Totals	292.4	265.8	99.9%

All data were normalized to Silicon mass = 28, with dust to gas ratio = 1.1 (see text).

Assuming a primitive dust-to-ice mass ratio of 1.10 (see the Dust-to-Ice Ratio section), we have an amount of $CO/H_2O = 0.07$ that will be released in an extended source by a two-step process:

1. Vaporization of the CHON grains or CHON mantles, compatible with a characteristic lifetime of about 2×10^4 seconds;

2. Photodissociation of formic acid and formaldehyde into CO plus other molecular fragments, with a shorter lifetime, about 5×10^3 seconds for H_2CO and 2×10^3 seconds for HCO-OH, from their cross sections and the solar flux.

The CO flux produced by the extended source, as observed by mass spectrometry of the neutral gas, is $(13 \pm 4)\% \times 0.6 = (8 \pm 2)\%$ of H_2, in reasonable agreement with the 7% deduced from the vaporization of the CHON grains.

We conclude that the release and photodissociation of formic acid and formaldehyde, which represent more than 5% by mass of the dust, can explain the total amount of CO released in an extended source of 15,000 km around the nucleus, as observed by the neutral mass spectrometer of Giotto.

We can generalize the previous approach and predict that the CHON grains are the source of other volatile species. A major one is hydrogen cyanide (HCN). HCN was already deduced to be the parent molecule of CN by quantitative arguments. Since it is present in the CHON grains, its slow release explains the persistence of the spiral-shaped CN jets observed in the coma of comet Halley. It can also be expected that CHON-type grains may play an important role in the exchange of matter between volatile dust and gas. During perihelion passage, their thermal evolution in the cometary mantle from volatile compounds (like formaldehyde) into stable dust (polyformaldehyde) could explain:

1. A slow evolution toward less gas and more dust (a larger and larger dust-to-ice mass ratio);

2. A slow transfer of carbon from gas to dust;

3. The disappearance of an excess source of carbon in the coma of evolved comets;

4. A reduction of the C/O ratio in the coma of evolved comets, depending on their perihelion distance and the number of their previous passages;

5. The claimed "puzzle" of the carbon abundance variation in the coma.

Table 10-4: Comparison of Elemental Abundances in Atom Numbers

Element	Solar	Comet Halley	Chondrites*		
			CI	CM	CV, CO
H	26600.0	30.0	2.0	1.0	0.10
C	11.7	8.0	0.7	0.4	0.08
N	2.3	1.4	0.05	0.04	0.01
O	18.4	18.8	7.5	5.3	4.10
S	0.5	0.6	0.5	0.2	0.12
Si	1.0	1.0	1.0	1.0	1.00

*Rounded averages.

Elemental Balance Sheet of Comet Halley

To establish the elemental balance sheet of comet Halley, the gas phase and the solid phase must be combined in the proper ratios. The elemental ratios of the gas phase, given in table 10-1, result from a detailed discussion of the numerous data published from 1986 to 1988, which is needed to resolve the numerous inconsistencies found in the literature. The error bars of table 10-1 imply a value judgment about these inconsistencies. The whole discussion will not be repeated here (see Delsemme 1991 in Additional Reading for further details).

The elemental ratios for cometary dust are those given from the sum of 80 dust particles that hit the mass spectrometer on board the Vega 1 Spacecraft.

Table 10-2 combines gas and dust by normalizing the amount of gas by the dust-to-gas mass ratio of 1.10. This ratio is confirmed by the fact that the dust composition is quasi-chondritic, but contains 3/4 of the total carbon of the comet, itself representing (when normalization to 1.10 is used) a nominal 3/4 of the cosmic carbon abundance.

Table 10-4 compares the elemental abundances in comet Halley from table 10-2 (expressed this time in atom numbers) with the solar and chondritic abundances. It is now quite clear that, except for hydrogen, depleted by a factor of almost 500, all the abundances of the light elements in comet Halley are within the error bars of solar abundances. The slight carbon deficiency may come from the variable CO^+ fraction blown away by the solar wind, hence not counted in its entirety in the atomic carbon line of the International Ultraviolet Explorer. The big carbon deficiency reported for other comets has disappeared entirely, mainly because of the large amount of carbon now detected in the dust. In the same way, the slight nitrogen deficiency may come from the N_2^+ depletion by dissociative recombination with electrons due to the magnetic confinement of the plasma in the onset of the tail. Undetected nitrogen could be blown away at high velocity in the plasma tail by this mechanism.

Quantitative Molecular Analysis of the Volatile Fraction

A useful tool to find discrepancies in the amount of the different molecular species detected by distinct methods (UV, visual, IR, or mass spectroscopy) is to build the balance sheet for all the elements of the molecules that have been observed or deduced, and verify that their sums reproduce (within the error bars) the elemental ratios observed or deduced in table 10-3.

The author has published a heuristic model of the chemical composition of Halley's volatile fraction, based on this technique. The model is reproduced in table 10-5. Each line of table 10-5 represents the contribution of each molecule to the elemental abundances. The total of each column gives therefore a number proportional to the total abundance of a particular element. The ratios of the totals must be compared with the elemental abundance ratios independently established and given in table 10-3.

The building of such a balance sheet for the elements remains difficult because of the large number of simultaneous constraints which are sometimes self contradictory. For instance, the known production rates for water and for oxygen-bearing compounds are in contradiction with the Pioneer Venus data that give $H/O = 1.4$. It is impossible to achieve an H/O ratio smaller than 1.8 (the ratio adopted in table 10-2) and it remains a feat to come close to it in table 10-4, because most other molecules add at least two hydrogen atoms per extra oxygen atom.

The only three molecules that are the exception are CO, CO_2, and H_2CO_2. However, the ratio CO_2/H_2O has been measured and cannot be changed beyond its error bars. H_2CO_2 (formic acid) and H_2CO (formaldehyde), discovered in the organic fraction of the dust grains, are volatile enough to be counted in the volatile fraction: formaldehyde is much more volatile than water, and formic acid has almost exactly the same vapor pressure as water (it

Table 10-5: Chemical Composition of the Volatile Fraction of Comet Halley

Parent	%	H	C	O	N	S
H_2O	80	160	—	80	—	—
H_2CO_2	4.5	9	4.5	9	—	—
H_2CO	4.0	8	4	4	—	—
CO_2	3.5	—	3.5	7	—	—
CO	1.5	—	1.5	1.5	—	—
CH_4	1.0	4	1	—	—	—
C_2H_2	1.5	3	3	—	—	—
C_3H_2	0.2	0.4	0.6	—	—	—
N_2H_4	0.2	0.8	—	—	0.4	—
HCN	1.0	1.0	1	—	1	—
N_2	0.5	—	—	—	1	—
$H_4C_5N_4$	0.5	2	2.5	—	2	—
NH_3	1.0	3	—	—	1	—
S_2	0.2	—	—	—	—	0.4
H_2S	0.2	0.4	—	—	—	0.2
CS_2	0.2	—	0.2	—	—	0.4
Totals:	100	191.6	21.8	101.5	5.4	1.0

Ratios:	H/O	C/O	N/O	S/O
from table 10-2:	1.8 ± 0.4	0.20 ± 0.05	0.10 ± 0.05	0.01 ± 0.05
from this table:	1.89	0.21	0.05	0.01

boils at 101°C). Both have been assumed to be the parent molecules for the extended source of CO that has been observed, since both photodissociate in times shorter than the vaporization of grain clusters. However, a fraction of this formaldehyde may be in polyformaldehyde; it has not been counted separately in table 10-4. An amount of formic acid somewhat larger than 4.5% would be beneficial to the balance of table 10-4, because it would improve the D/H ratio.

The total production of CO has been set at 10% of the volatiles, that is, 12.5% of the production of water; CO_2 is set at 3.5% of the volatiles (or 4.2% of water). The total production rate of CO and CO_2 has been pushed to the maximum compatible with the error bars of the observations to keep both H/O and C/O within reasonable limits.

Methane (CH_4) and ammonia (NH_3) represent altogether about 3 to 4% of water. Acetylene C_2H_2 is assumed to be the source of C_2 and C_3H_2 the source of C_3; C_3H_2 has now been shown to be abundant in interstellar space. Hydrogen cyanide is the source of CN. Finally, the "normal" ratios C_2: C_3: CN = 1.5 : 0.2 : 1 have been adopted here.

The amount of sulfur molecules is small enough not to interfere very much with the H and the C balance sheet. A more serious problem might turn out in the nitrogen balance. Hydrazine (N_2H_4) is the probable source of NH_2, whereas NH_3 does not seem to be very abundant: a bonus for the balance sheet because NH_3 would introduce much hydrogen and no oxygen. It is also difficult to introduce more than 1% of N_2 as suggested by the small production rate of N_2^+. A recent result reduces N_2 to a welcome 0.5%.

It is interesting to note that neither N_2 nor CO can be very abundant in a nucleus whose temperature is driven by the vaporization of water ice; this is because of the very large vapor pressure of N_2 and of CO. Even their clathrates have high dissociation pressures. Incidentally, the assumption that carbon monoxide can form a clathrate with water has been recently confirmed. It has a dissociation pressure of 1256 bars at 273 K and it lies, therefore, between the oxygen and the nitrogen clathrates in stability.

The observed presence of purine ($H_4C_5N_4$) in the organic fraction of the dust, must not be taken at its face value in table 10-5. It has been used here as a symbol and a reminder that we may still have several unknown organic compounds in the dust that might dissociate or vaporize into the coma.

Altogether, our understanding of the volatile fraction has considerably improved with comet Halley. The major recent achievement is, of course, the understanding that a very large fraction of the dust is volatile enough to be the origin of an extended source of gas in the inner coma. However, most of the exobiology questions remain pending. In particular, those about purines and pyrimidines must wait for a rendezvous mission to a comet.

Isotopic Ratios

Before comet Halley, the only isotopic ratio approximately measured in comets was $^{12}C/^{13}C$, but the error bars were so large that it was impossible to say whether it was really different from 89, the terrestrial ratio.

With the flybys of comet Halley, new isotopic ratios have been measured. It has been reported that the Giotto neutral mass spectrometer gives a D/H ratio between 60 and 480 ppm, implying that the cometary deuterium is enriched by a factor of at least 3 with respect to interstellar hydrogen, or to hydrogen in Jupiter and Saturn. A comparable enrichment occurs in terrestrial water, suggesting that water on the terrestrial planets can be of cometary origin, an idea that was proposed earlier by models of the accretion of the planets from the solar nebula, and that is described in detail in the second part of this chapter.

It also has been reported that $^{18}O/^{16}O = (23 \pm 6)10^{-4}$ in comet Halley, agreeing within the error limits, with the average solar system value of 20×10^{-4}. Although we still lack much data, nothing has been found yet that would be inconsistent with cometary material originating from the same reservoir as the rest of the solar system.

Exogenous Origin of Carbon on Terrestrial Planets

Another interface between comets and exobiology cannot be neglected here; it concerns the exogenous origin of carbon on the terrestrial planets and, in particular, the origin of what became the biosphere on Earth.

The origin of carbon on the terrestrial planets is not trivial because there is mounting evidence that, at a time when the largest planetary bodies were micrometer-sized grains of dust, there was a high-temperature period in the zones of the future terrestrial planets; hence all dust was outgassed and most carbon was stored as CO in the gas phase just before planetesimals formed.

Evidence for a High-Temperature Phase

The formation of single stars generally implies the transient existence of an accretion disk. Such a disk is needed to get rid of the excess angular momentum during the collapse of an interstellar cloud nodule. The disk stops the free-fall collapse of the cloud; viscous friction brakes its fast-turning inner layers and accelerates the more slowly turning outer layers. As a result, mass is transported inward and feeds the central star, whereas the excess angular momentum is transported outward and widens the disk.

The theoretical analysis has been recently supported by the discovery of many dust disks surrounding young stars. First, the Infrared Astronomical Satellite (IRAS) detected large infrared excesses near 100 micrometers in the spectra of many young stars. Their interpretation as dust disks was confirmed in particular by a photograph of the dust disk of Beta Pictoris.

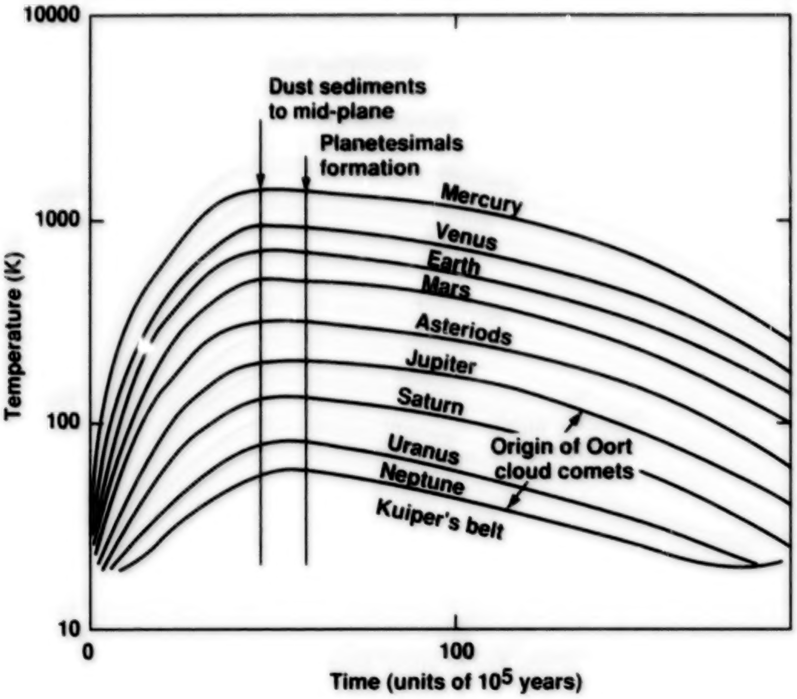

Figure 10-1. Typical example of the variation of the gas temperature in the accretion disk in the various regions of planetary formation.

A common feature of the different models of viscous accretion disks is that the temperature dependence on heliocentric distance r follows a law close to $r^{-3/4}$, regardless of the details of the viscous mechanism. However, this temperature varies with time (see fig. 10-1): it grows steadily during the collapse of the interstellar cloud that feeds the disk and starts falling as soon as the accretion subsides and stops. Typically, the temperature varies with $(M\dot{M})^{1/4}$, M being the total mass (disk plus star) and $\dot{M}$ the mass accretion rate. For reasonable values of the parameters, for instance $M = 1M_0$ and $\dot{M} = 10^{-5}\ M_0\ yr^{-1}$, the maximum temperature reached at each heliocentric distance is reasonably consistent with those inferred from the chemical constitution of the terrestrial planets. The temperature gradient inferred is also consistent with the $r^{-3/4}$ law deduced for simple viscous disks. More sophisticated models give a slightly different temperature gradient for the mid-plane temperature (for instance in $r^{-9/10}$), giving an even better agreement.

In this model, the existence of reduced iron in the nucleus of the Earth is explained by the reduction of iron in the fine dust grains of the disk which are in contact with a large amount of H_2 at a temperature close to 900 K. Iron must be reduced before the accretion of dust into large objects; there is no way to reduce iron after it has reached the Earth's mantle in the form of an oxide or silicate.

Using such a high temperature for the Earth's zone of accretion, the $r^{-3/4}$ dependence on distance predicts 450 K near 2.6 AU, which is the place beyond which the C asteroids start to outnumber the S asteroids. There is a strong suggestion (from infrared spectra and very low albedos) that carbonaceous chondrites come from the C asteroids, and 450 K is the temperature required to separate the formation of ordinary from carbonaceous chondrites. These two examples are sufficient to illustrate the type of indirect arguments that can be used to support this fossil evidence, but there are several others that result in the same conclusions.

Now let's consider models of the accretion disk when it reaches its maximum temperature. Figure 10-2 illustrates three adiabats for the gas in the mid-plane of the accretion disk at that moment. These adiabats have been traced for illustrative purposes only. The symbols for the planets are located in the center of their accretion zones. To satisfy the previous temperature requirements, model C has too much mass and model D not enough. Model CD has been selected by interpolating models C and D; it fits in with the previously derived formation temperature of the planets. The three adiabats have been superimposed on a thermochemical diagram giving the different domains of the major molecular species for a gas of solar composition. The three adiabats show that, in the zones of the terrestrial planets, carbon is in the form of gaseous CO. The three adiabats have the advantage of showing the extreme range that would remain consistent with planetary data, like for model D the reduction of the Earth's core, or for model C the existence of carbonaceous chondrites at 2.6 AU in the asteroid belt.

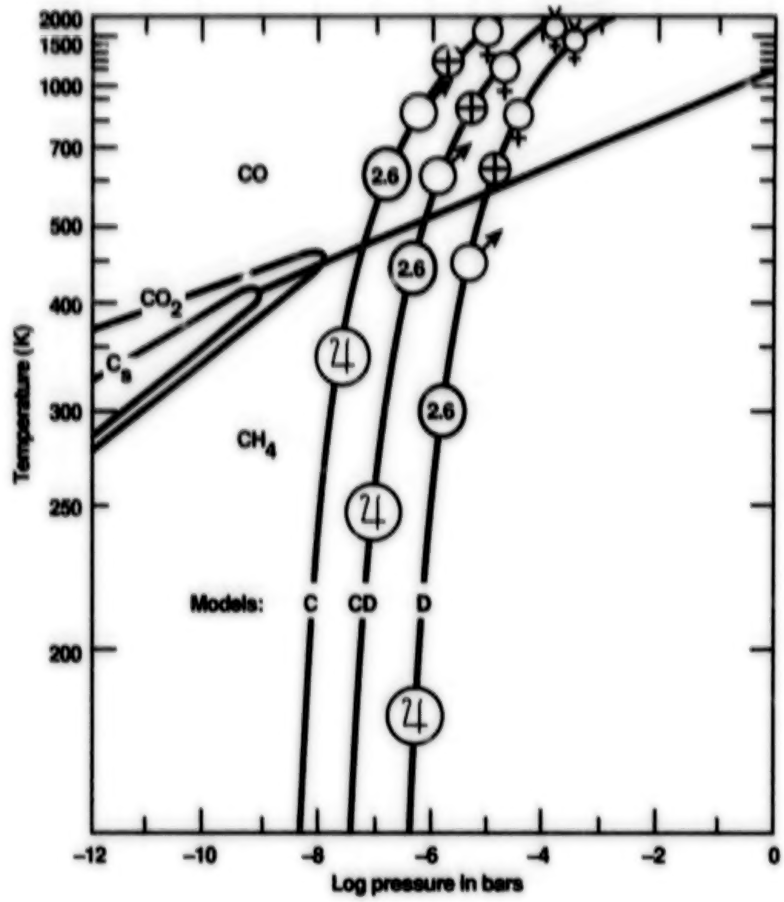

Figure 10-2. The thermochemical equilibrium diagram of the carbon compounds in a gas of solar composition. The three quasi-vertical curves represent three adiabats through the mid-plane of the accretion disk at the time of its maximum temperature. Models C and D correspond to solar-nebula models C and D; model CD is a linear interpolation of the two previous models that fits in better with an empirical adiabat derived from the planets' chemical composition. The other solid lines separate the domains in which CH_4, CO, CO_2, and graphite (C_s) are the most abundant carbon compounds. The activity of graphite is a maximum not far from the line separating CO and CH_4. All adiabats compatible with planetary chemistry show that the terrestrial planets' formation took place in a hot portion of the accretion disk, where a large hydrogen excess reduced and dehydrated all dust grains. The symbols of the planets locate their zones of formation. The figure 2.6 in a circle refers to 2.6 AU, beyond which the C asteroids begin to outnumber the S asteroids. The infrared spectra and dark albedo of the C asteroids suggest carbonaceous chondritic material, hence the activity of graphite should be near maximum a bit beyond 2.6 AU. This implies that the grains of the Earth's zone were too hot to contain enough carbon or water. The bulk of our oceans and of our carbonates must therefore have an exogenous origin.

To establish whether thermochemical equilibrium is reached, we must compare the durations available in the accretion disk with the time constants of the chemical kinetics. The durations in the disk are given by the models. In particular, the trend toward equilibrium will begin only when large-scale turbulence subsides and stops. At that time, several quasi simultaneous events occur quickly (see fig. 10-1): the maximum temperature is reached; the Sun's mass stops growing; and the dust sediments and forms thin dusty rings in the mid-plane, where it starts accreting into larger solid objects (the planetesimals). This is a very effective way to separate solids from gas before the temperature drops. A duration of 10^3 to 10^4 years is available to reach thermochemical equilibrium in the disk. The end of that duration occurs as soon as solids are effectively removed from the gas, that is, when dust has agglomerated into boulders of at least a few meters in diameter.

Lewis and co-workers have examined the chemical kinetics of a gas of solar composition. The largest chemical time constants that are significant turn out to be of the order of one century near 1000 K; therefore thermochemical equilibrium will be reached before solids are separated from the gas phase. It is concluded that the dust has been completely outgassed, dehydrated, and reduced in the zone of the terrestrial planets, before its incorporation into planetesimals, and eventually into planetary bodies.

These researchers have also carefully considered the carbon problem, and they have been puzzled by the retention of carbon in the terrestrial planets. Searching for a mechanism of carbon retention for the terrestrial planets, Lewis and co-workers have deduced that the only way to imprison carbon in the solid phase is to put it in solution in metallic iron grains, but its amount would still be two or three times smaller than the observed amount of carbon on Earth and Venus.

Their choice for the adiabat has, however, put the Earth's zone near the peak of graphite activity; for any adiabat, graphite activity goes through a maximum in the vicinity of the line separating, on figure 10-2, the domains of CH_4 and CO. Adiabat CD in figure 10-2 puts the formation temperature of the carbonaceous chondrites (assumed to form at 2.6 AU and beyond) from 300 to 450 K in agreement with other assessments, and close to the maximum of graphite activity, favorable for Fischer-Tropsch type reactions. But this removes the Earth's zone of accretion from the maximum of graphite activity; the amount of carbon available for the Earth and Venus becomes several hundred times smaller than the observed amount on these planets. Hence the interpretation of the organic compounds in carbonaceous chondrites as the result of Fischer-Tropsch-type reactions strongly implies that the bulk of the observed carbon compounds on the terrestrial planets has an exogenic origin.

From Planetesimals to Planets

After accretion of planetesimals, the pace slows down drastically. Larger planetary embryos grow from numerous collisions. It has been established that collisional velocities grow in proportion to the size of the largest objects. The reason is that the growing gravitation of the embryos deflects the smaller bodies into more elliptical orbits. This means that the zone feeding a protoplanet becomes wider and wider; objects deflected by the giant protoplanets may come closer and may eventually hit the Earth. Since these objects come from colder zones (see fig. 10-1) they contain snow and organic matter; hence the final steps of accretion become heterogeneous.

A Veneer of Volatile and Organic Material

Silicaceous planetesimals originating in the zones of the terrestrial planets were probably considerably depleted when these planets reached 90% of their masses. Since the accretion of the giant planets took longer (at least for Uranus and Neptune) the diffusion of the planetesimals coming from larger and larger distances took longer and longer, implying that the ordinary chondrites arrived first (say from 1.6 to 3.6 AU, possibly bringing 7% of the final mass); the carbonaceous chondrites arrived later (possibly 2% of the final mass, from 2.6 to 5 AU); the comets arrived last, bringing 1% of the final mass from 5 to 50 AU and beyond. Oxidized material being added as a late veneer achieved the high oxidation state of the Earth's crust.

Cometary Contribution

Using a model from 1972, the author assessed that the total mass of comets hitting the Earth at the end of accretion was large enough to explain all the water in the oceans and all the carbonates and organic matter on the Earth (from table 10-5, in round figures, water represents 40% of the cometary mass and organic matter more than 20%).

Better evaluations have now been made. Others have found that comets have brought four times as much water as the mass of our oceans. Still other researchers have found ten times the present mass of water. It has also been determined that the total mass of water in the oceans is explained if only 10% of the extant lunar craters are due to comets. The present amount of water and carbon on the Earth is consistent with a considerable erosion of the atmosphere and oceans, due to the last giant impacts at the end of the accretion period. These giant impacts are predicted by the accretion theory, and their remnants are visible everywhere (Hellas Planitia on Mars, Mare Imbrium on the Moon, Caloris basin on Mercury, etc.). Giant impacts probably explain the origin of the Moon, as well as the considerable erosion of the atmosphere on Mars because of its small gravity.

The amount of organic material brought down to Earth during the accretion period (up to some 3.8 billion years ago) was somewhat larger than half the mass of the oceans, although the most volatile part may have been considerably eroded by giant impacts. Only an extremely small fraction of such a mass needs to be preserved in the form of prebiotic compounds to completely overthrow the traditional picture in which the first prebiotic syntheses occurred on Earth.

Concluding Remarks

The transfer of organic compounds from interstellar space to the outskirts of a protoplanetary disk, their accretion into cometary objects, and the transport of the latter into the inner solar system by orbital diffusion throw a new light on the central problem of exobiology. It suggests the existence of a cosmic mechanism, working everywhere, that can supply prebiotic compounds to ubiquitous rocky planets, in search of the proper environment (proper redox ratio, liquid water, etc.) to start life in many places in the Universe.

However, we do not know enough about cometary chemistry to answer most of the exobiology questions. The multiple flybys close to the nucleus of comet Halley have brought a wealth of new information on comets, in general, and have come tantalizingly close to answering important exobiology questions. Nonetheless, the answers have been left in doubt mostly because the fast flyby velocity destroyed significant molecules.

We need other cometary missions if we want to answer these exobiology questions. The first prerequisite is a rendezvous mission, that is, a mission whose relative velocity to the cometary nucleus can be brought down to a few meters per second. In such a mission, we also need a penetrator to reach pristine material a few feet under the surface layer that has been modified by irradiation and vaporization. Later on, it is clear that only a sample return mission will be able to answer the final questions, but such a mission cannot be prepared in haste before having studied a cometary nucleus from nearby. A rendezvous mission is therefore a prerequisite for clarifying sampling techniques.

The choice of a proper comet is also an important consideration. Until recently, short-period comets were considered as a last resort because they are the only ones whose orbits are known sufficiently ahead of time to prepare a mission, but the "new" comets (coming straight from the Oort cloud, where they had been kept in the deep cold of space) were considered more "pristine" than short-period comets. Recently however, it has been established that the orbital diffusion of comets cannot produce the flattened symmetry of the system of short-period comets, if their source does not already have the same flattened symmetry. It has been deduced that the source of the short-period comets is not the (quasi-spherical) Oort cloud, but the Kuiper belt proposed in 1951. The Kuiper belt would be a remnant of the accretion disk (like the one around Beta Pictoris) where cometesimals never formed planets. Figure 10-1 shows that the maximum temperature in the Kuiper belt never exceeded 40 K, whereas other mechanisms put the source of the Oort cloud comets in the zones of the giant planets (the diffusion of their orbits is due to the growing embryos of the giant planets), where the temperature reached from 50 to 180 K. Paradoxically, the short-period comets, at least before entering the inner solar system, could be more pristine objects than the Oort cloud comets (this giant reservoir from which all long-period comets originate). Short-period comets that have entered the inner solar system only very recently and which have perihelion distances as large as possible represent the best targets of opportunity for exobiology studies.

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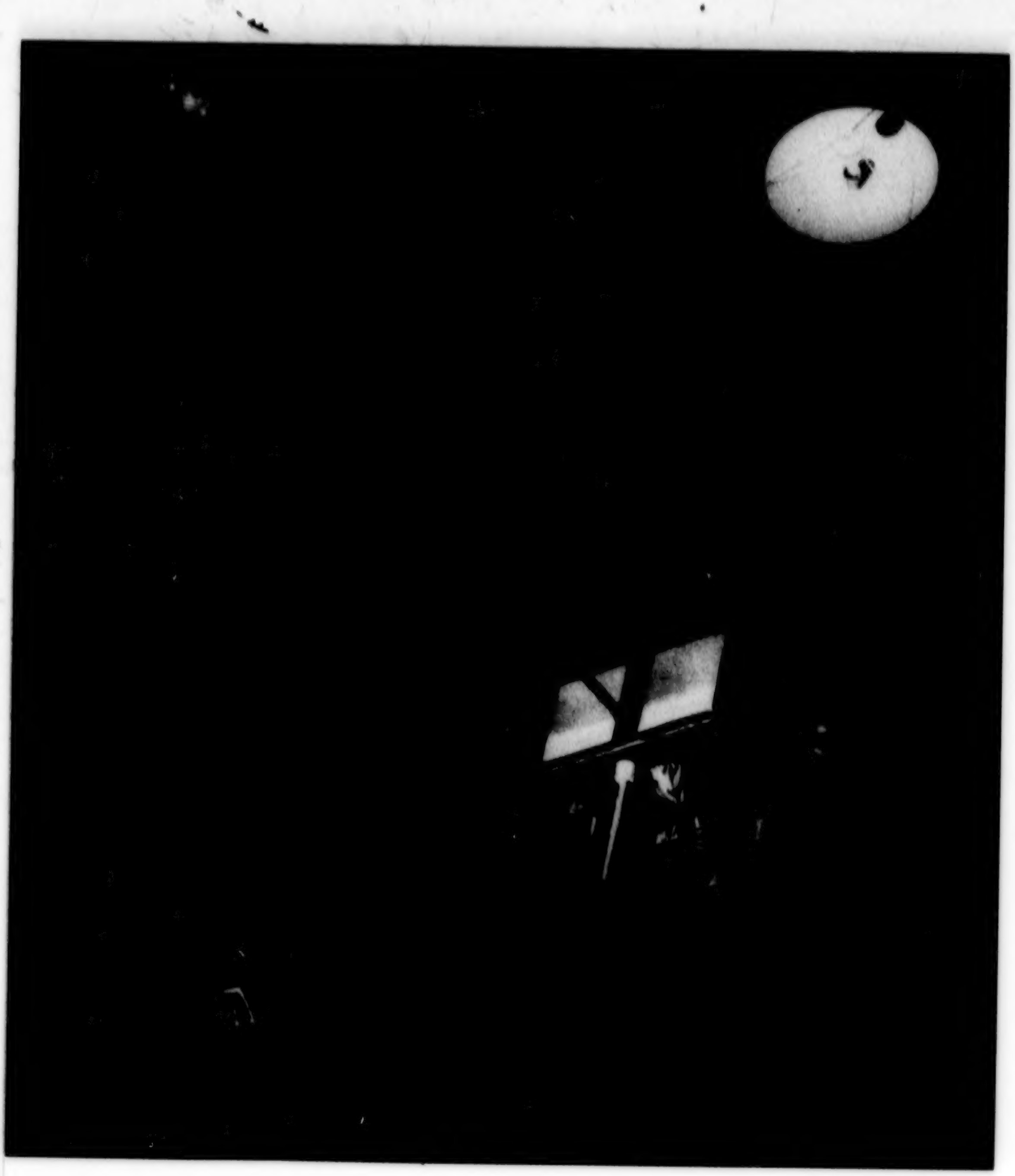

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Mars Observer: Mission Toward a Basic Understanding of Mars

A. L. Albee

The Mars Observer Mission will provide a spacecraft platform about Mars from which the entire Martian surface and atmosphere will be observed and mapped by remote sensing instruments for at least 1 Martian year. The scientific objectives for the Mission emphasize qualitative and quantitative determination of the elemental and mineralogical composition of the surface; measurement of the global surface topography, gravity field, and magnetic field; and the development of a synoptic data base of climatological conditions. The Mission will provide a basic global understanding of Mars as it exists today and will provide a framework for understanding its past.

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This chapter describes the Mars Observer Mission as it stood in the late 1980s. Since that time, some changes have occurred in the mission plan. Mars Observer will be launched on a Titan III and will have a somewhat higher mapping orbit and somewhat different mapping cycle. The Visual and Infrared Mapping Spectrometer experiment will not be flown and the radar altimeter has been replaced by a laser altimeter. For an updated description see Albee, A. L. and Palluconi, D. F. (1990) EOS, vol. 71, pp. 1099, 1107.

Science Objectives and Expected Results

The formal scientific objectives of this geoscience and climatological mission are to

1. Determine the global elemental and mineralogical character of the surface material;

2. Define globally the topography and gravitational field;

3. Establish the nature of the magnetic field;

4. Determine the time and space distribution, abundance, sources, and sinks of volatile material and dust over a seasonal cycle;

5. Explore the structure and aspects of the circulation of the atmosphere.

These first-order scientific objectives can be addressed within the framework of a low-cost orbital mission. All five objectives involve global mapping. For the geoscience objectives, this mapping is time-independent and therefore two-dimensional: latitude and longitude. For many climatology objectives, the mapping is four-dimensional: latitude, longitude, vertical, and season. As a result of the Mission we should have a more systematic global characterization of Mars, as it exists today, than of Earth. This characterization will help us to understand the geologic and climatologic history of Mars and the evolution of its interior and surface.

Geoscience

The first three of these objectives are geoscience objectives—they involve measurement of quantities of geological, geochemical, and geophysical interest. The single lithospheric plate comprising the Martian surface is a little larger than the combined area of all the Earth's continental plates. Mapping of this 144×10^6 km^2 area will occur over the full Martian year—687 Earth days—planned for the Mars Observer Mission. This plate is highly complex with high-altitude, heavily cratered uplands in the southern hemisphere and lower, less-cratered plains in the northern hemisphere. Relatively young volcanic shields and cones are most abundant in the north, but annual permanent ice and associated layered deposits occur near each pole. Many units have been variously modified by crustal and surface processes to produce canyons and channels; chaotic, fretted, and hummocky terrains; and sand sheets and dunes.

The Mission will seek to understand the distribution of chemical elements and minerals on the Martian surface in relation to the age, origin, nature, and weathering of the surface rocks. Simultaneous global mapping of the gravitational field, the surface topography, and the magnetic field will improve our understanding of both the surface and the interior of Mars. Combining these distinctly different measurements in a single mission exploits their inherent synergism and allows us to realistically address such global problems as Mars' bulk composition, the degree of differentiation of the planet, the chemical and mineralogical composition of the crust and mantle, and the nature and cause of the planet-wide dichotomy into uplands and plains.

The potassium/uranium ratio, for example, provides important clues. Uranium provides an index for those elements that tend to condense from the solar nebula at high temperatures, potassium is an index for elements with lower temperatures, yet both behave similarly in magmatic processes. Thus the K/U ratio, coupled with other elemental ratios, provides a means of reconstructing the bulk composition of the planet. The absolute content of these two elements in the crustal material also gives an indication of the degree of differentiation of the planet; gives an estimate of the amount of volatiles, including water, that has outgassed from the planet; and places tight constraints on the thermal modeling of the planet.

The nature of the crust will be well defined by the elemental and mineralogical mapping, and global variations in crustal thickness will be determined from mapping the gravity field and surface topography. Major unknowns are the nature and cause of the planet-wide dichotomy into old cratered upland and sparsely cratered plains. Mars Observer may determine chemical and mineralogical differences between the two regions, how the crust and lithosphere differ beneath the two regions, and the nature of their boundary. All of these will lead to a much better understanding of the nature of the dichotomy, what caused it, and when it developed.

The volcanoes of Mars are well known, but we have almost no information on the chemistry and mineralogy of the lavas—from which we could infer the composition of the mantle, the depth of origin of the magma, and whether the magmas have undergone differentiation during their passage from the source region to the surface. Mars Observer should discover whether there are regional differences in composition that might be the result of differences in source rock or depth of origin of the magmas.

Mapping of the gravitational and topographic fields will lead to better understanding of the volcanic processes and the thermal evolution of the interior. Lithospheric thickness may be determinable from the deformation of the surface by large volcanoes. Thermal conditions in the interior can be understood from the apparent viscosity of the crust and upper mantle as indicated by the degree and depth of isostatic compensation of different craters, canyons, and volcanoes. Increase in the depth of composition with the age of such features might suggest that the lithosphere has thickened with time. Magnetic measurements will indicate whether conditions in the core today can sustain a planetary dynamo and, if not, whether such conditions existed in the past.

Much of the Martian surface appears to be covered with weathering products. Tentative identifications have been made of specific absorption bands seen in telescopic observations, but we actually know very little about the chemistry and mineralogy of the surface. Mars Observer will provide these basic characterizations. We particularly want to understand the amount of water, how it is contained in the minerals, and whether these minerals are forming currently or are "fossils" from past climates. We may then infer how water is cycled during the current seasonal and obliquity cycles and whether significantly different climates in the past are required to explain the formation of the soil materials. Understanding the abundance of volatile-containing materials will lead to improved estimates of the amount of water and other volatiles outgassed from the planet and provide clues as to when the outgassed volatiles were removed from the atmosphere to become fixed in the soil.

Mars Observer provides an opportunity for mapping the planet-wide distribution of ice in the near-surface material. Ice is believed to be a major component in the polar-layered terrain, detectable by Mars Observer from both (spectral) composition and gravity information. Ice may also be present in the soils at high latitudes, its low latitude limit varying with the season. Detection of such ice will depend upon its proximity to the surface and whether seasonal changes can be discerned. Detection of liquid water is not expected, but aberrant conditions might allow seepage to the surface. Such a discovery would lead to significant reassessment of how water is circulated about Mars.

Finally, more precise determination of topography and gravity will lead to a more complete understanding of almost every geologic process that has affected the surface. Our current understandings are based largely on surface morphology and are hindered by the lack of quantitative information against which different ideas can be tested. Better information on slopes will lead to a better understanding of fluvial erosion and emplacement of lavas. Comparison of present-day slopes with past slopes as indicated by lava flow or stream directions will provide information on deformation rates. Better information on gravity patterns of craters and on crater depths, rim heights, and ejecta thickness will lead to a more complete understanding of cratering.

Acquisition of chemical, mineralogical, gravitational, and elevational data by Mars Observer will improve our understanding of almost every aspect of the planet's geology. However, the main impact will be an improved understanding of global problems rather than local geologic processes. Improved understanding of the planet's bulk composition, the composition of the crust and mantle, the thickness of the lithosphere, the thermal state of the interior, and the planet's outgassing history will inevitably follow from Mars Observer. This will be the Mission's main legacy.

Climate

The fourth and fifth objectives address volatile cycles (e.g., carbon dioxide, water), dust, the Martian atmosphere, and the atmospheric interaction with the surface. The emphasis is on the seasonal variations, hence the use of the term "climatology." An understanding of the current climate of Mars permits us to better assess how currently active processes (weathering, erosion, atmospheric transport, dust deposition) are modifying the surface. If we are able to understand the current climate, we may more confidently project this knowledge backward in time to periods when the Martian orbit, axial characteristics, and atmospheric pressure were different. In principle, determination of the daily, as well as the seasonal, behavior of the global atmosphere would be ideal. However, providing complete daily sampling is difficult from a single orbiting spacecraft and is not possible from the sun-synchronous mapping orbit selected for this Mission.

Mars Observer will provide

1. A greatly improved characterization of the atmospheric general circulation of Mars and of the factors that govern it;

2. A much more complete description of the present seasonal cycles of H_2O, CO_2, and dust, and the key parameters controlling them;

3. Important clues and constraints on the nature of past climate changes; and

4. The evolution of the Martian atmosphere.

The good synoptic coverage of the atmospheric temperature will make it possible to assess the relative importance of various circulation mechanisms in transporting heat and momentum as well as in transporting volatiles and dust. By conducting observations over a full set of Martian seasons and at times of varied dust loading and cloud cover, it should be possible to identify the feedback between the circulation, the volatiles, and the dust—how each affects the other, and how they collectively affect the climate.

These data and their analysis will provide a solid basis for comparing the general circulation of the Martian and terrestrial atmospheres. By using a variety of dynamic models to seek out their similarities and differences we will gain a more profound perspective on atmospheric dynamics and its underlying controls.

Mars Observer will provide quantitative information on almost all major fluxes and reservoirs that control the seasonal cycles of CO_2, H_2O, and dust. For example, these data will help define the amount of atmospheric water gained or lost at different seasons from the principal surface water reservoirs—the seasonal polar caps, the perennial polar caps, and the mid- to low-latitude regolith. By measuring the amount of atmospheric water present, both as vapor and as ice clouds, and by concurrently obtaining data on atmospheric motions, Mars Observer will permit an evaluation of how far atmospheric water travels from its source region. Thus the relative roles played in the seasonal water cycle by the polar caps, the regolith, and atmospheric transport will be quantitatively assessed.

Mars Observer will also provide valuable insights into the factors controlling the life cycle of global dust storms. By providing good temporal and spatial coverage of the growth phase of global dust storms, by observing the atmospheric circulation, and by concurrently measuring the spatial distribution of atmospheric dust and temperature, a firm basis will be provided for understanding the processes that permit a local dust storm to grow rapidly to global proportions and then to decay soon after. Of special interest will be determination of the dust heating/wind speed relationships and understanding the influence of dust loading over the life span of a global dust storm.

It will also be possible to examine the important interactions of the various seasonal cycles. For example, the amount of dust deposited on the seasonal and perennial polar ice deposits probably plays a critical role in the stability of these deposits. The occurrence of a perennial H_2O ice cap in the north and the occasional occurrence of a perennial CO_2 ice cap in the south may be due to differing amounts of dust deposited on the two caps, thereby strongly influencing their albedos. By observing the amount of dust present in the polar atmospheres over the course of a Martian year and by measuring the heat balance of the polar regions, the experiments on Mars Observer will provide an understanding of the coupling of the dust and CO_2 cycles.

Improved understanding of atmospheric circulation and seasonal cycles may provide insight into how they may have changed in the past due to astronomical variations of orbital and axial properties. This goal will be significantly enhanced if an extended Mars Observer Mission or future mission provides observations over several Martian years. We need to characterize the interannual variation in atmospheric circulation and the H_2O/CO_2/dust cycles in order to understand their causes.

Observations of the polar layered terrain will advance our knowledge of the nature of the quasi-periodic changes on Mars. Measurements of the dust:water:ice ratio, topography, and volume of these layers will provide constraints on their formation and the nature of the episodic atmospheric transport of water and fine dust into the polar regions. Determination of the amount and size of dust and water ice particles in the present polar atmosphere will provide an estimate of the present rate of deposition.

The occurrence, abundance, and spatial distribution of hydrous and carbonate minerals in the Martian soil is important to determining when and how H_2O and CO_2 have been lost from the atmosphere by weathering processes. These provide constraints on the possible occurrence of warmer and wetter climates in Mars' past history and on the total amount of H_2O that was outgassed into the atmosphere. The nature of fluvial channels provides additional constraints on the past climate.

With all of these observations and understandings it will be possible to explore the linkages between astronomical (primarily solar) variations, climate changes, and genesis of layered terrain and fluvial channels. Finally, given such improved understanding of the coupled issue of climatic change and atmospheric evolution on Mars, it will be important to make comparisons with comparable phenomena on Earth, seeking out general patterns and controls and the factors responsible for the similarities and differences in these phenomena for the two planets.

Exobiology

Scientific evidence makes it difficult to sustain the notion that life exists on Mars today, but evidence does suggest that the climate on early Mars was quite different from that today and may have been more conducive to life. Dendritic valley systems indicate that liquid water once flowed on the Martian surface and we can infer that the Martian surface temperatures were considerably warmer and atmospheric pressures much higher than they are today. Life could have arisen on Mars during this early climate epoch. If so, some evidence might eventually be found within the large areas of the Martian surface that date back to this time.

The Mars Observer Mission is unlikely to find such evidence, but it will provide information to guide future missions to locations on the Martian surface optimal for this search. Moreover, it will help us to understand the climatic history of both Mars and Earth. The daily and seasonal weather cycles on Mars and Earth are very similar, but the Martian thin atmosphere, rapid heating and cooling of the surface, the abundance and distribution of water, the lack of oceans, the annual condensation cycle of carbon dioxide, and massive dust storms are critical differences. Since both planets show evidence of drastic climate changes in the past, common mechanisms might be responsible, which include solar luminosity changes, orbital variations, volcanic eruptions, asteroid impacts, etc. The improved understanding of climatic change, with consequent effects on biota and biologic processes, will be the most important contribution of Mars Observer to the science of exobiology.

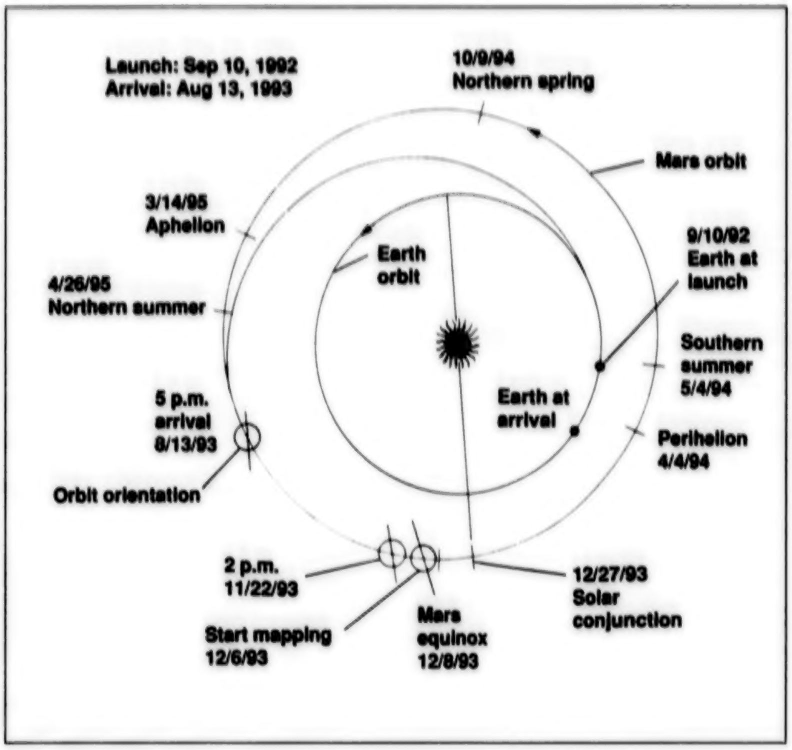

Launch: Sep 10, 1992
Arrival: Aug 13, 1993
10/9/94
Northern spring
Mars orbit
3/14/95
Aphelion
Earth
orbit
9/10/92
Earth at
launch
4/26/95
Northern summer
Southern
summer
5/4/94
5 p.m.
arrival
8/13/93
Earth at
arrival
Perihelion
4/4/94
Orbit orientation
2 p.m.
11/22/93
12/27/93
Solar
conjunction
Start mapping
12/6/93
Mars
equinox
12/8/93

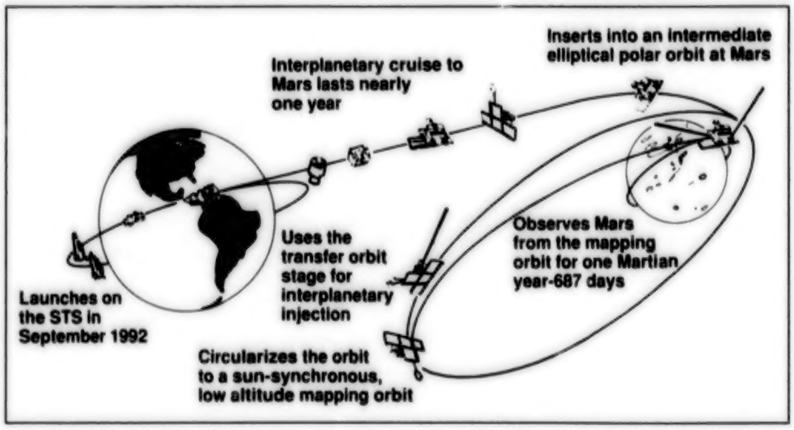
Figure 11-1. Mars Observer trajectory in heliocentric projection.

Mission Description

Mars Observer is planned for a September 1992 launch (fig. 11-1). After the 1-year transit the spacecraft is injected into an elliptical orbit about Mars with periopsis near the north pole (fig. 11-2). The orbit is then adjusted through a series of maneuvers to a near-circular, sun-synchronous (2:00 a.m./p.m.), low-altitude (360 km), near-polar orbit. During the Martian year in this mapping orbit the instruments gather data by repetitive global mapping. At the end of the Mission the spacecraft is boosted from the mapping orbit to a quarantine orbit in which it can remain indefinitely.

Figure 11-3 summarizes the mapping timeline, relative to Martian seasons and the normal dust storm period. The mapping orbit is a 2-hour orbit with a 3-day repeat cycle shifted by 30 km. As a result, the planet is repeatedly mapped in 59-day cycles with 40-km swath widths and 10-km overlaps. The Mars orbit insertion (MOI) period is shown ending just before solar conjunction and the beginning of the dust storm period. It is scientifically important to make observations, preferably for an entire 59-day cycle, before the onset of a major dust storm, and it is hoped that the fuel margin will permit a shorter MOI period. Figure 11-3 also shows that the playback data rate for an 8-hour link varies by a

Figure 11-2. Generalized Mission description. The launch may use a Titan III rather than the STS, as shown.


Inserts into an intermediate
elliptical polar orbit at Mars
Interplanetary cruise to
Mars lasts nearly
one year
Observes Mars
from the mapping
orbit for one Martian
year-687 days
Uses the
transfer orbit
stage for
interplanetary
injection
Launches on
the STS in
September 1992
Circularizes the orbit
to a sun-synchronous,
low altitude mapping orbit

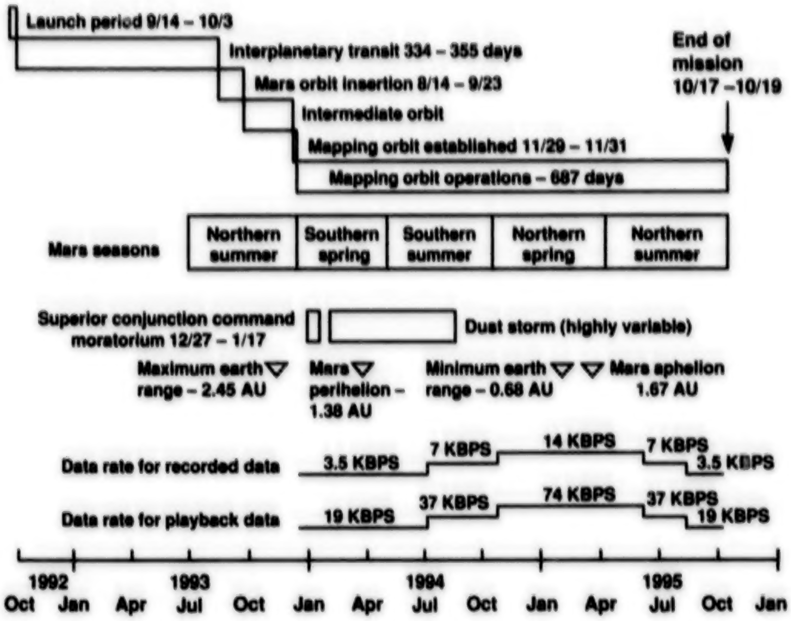


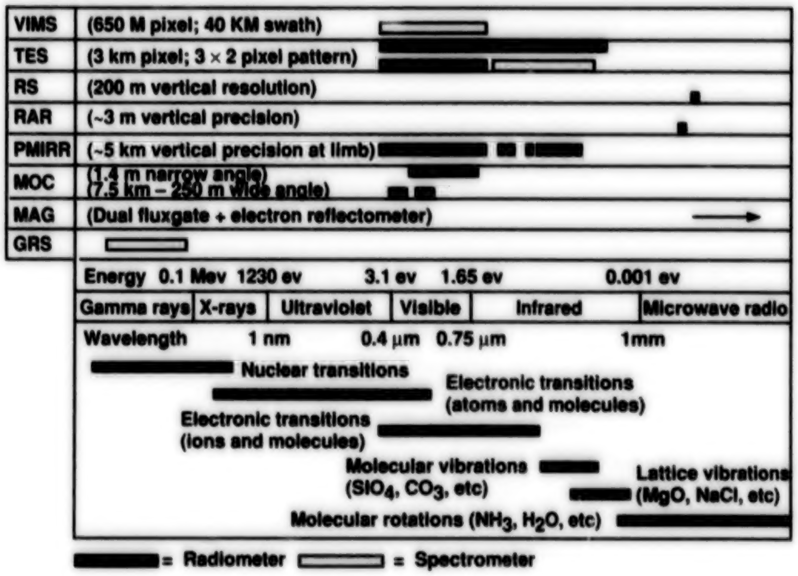
Figure 11-3. Mapping phase time line.

factor of four with Earth-Mars distance (see fig. 11-1) during the Mission. The continuous recorded data rate to the tape recorders is about one-fifth the playback rate.

Science Experiments and Instruments

The remote sensing instruments for Mars Observer have been chosen with the above stated scientific objectives in mind. Collectively the instruments cover much of the electromagnetic spectrum and, as shown in figure 11-4, they sense a variety of physical processes. Each instrument produces well-defined sets of measurements and addresses specific major objectives, but nearly every data set also contributes to a much wider variety of scientific investigations. Five interdisciplinary scientists have been selected in order to exploit the strong synergism that exists between the data from the eight instruments. Moreover, participating scientists, including 10 from the former USSR, will be added in the future to further exploit the data returned from the Mission. Each of the instruments and their experimental objectives are described in the next section.

Figure 11-4. Mars Observer instrument measurement ranges and physical processes of electromagnetic radiation.



The Gamma Ray Spectrometer and Magnetometer sensor assemblies are mounted on individual

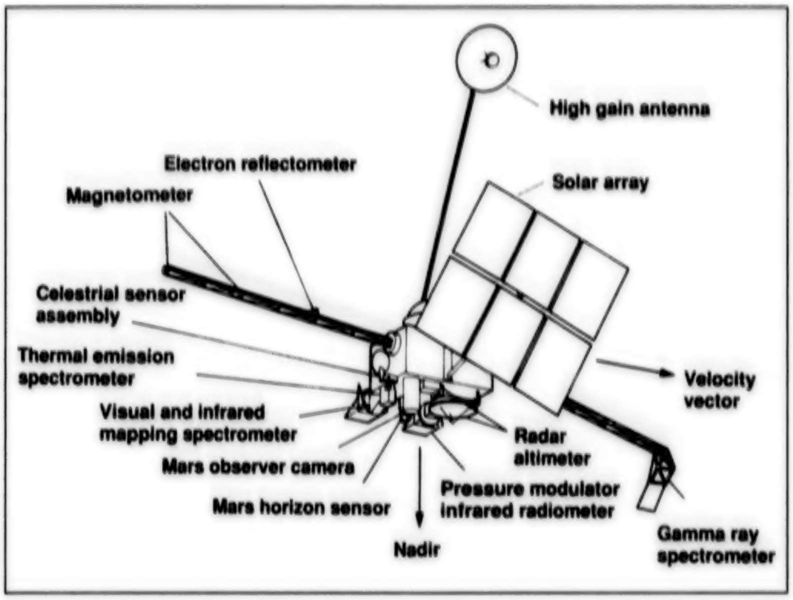

High gain antenna
Electron reflectometer
Magnetometer
Solar array
Celestrial sensor assembly
Thermal emission spectrometer
Velocity vector
Visual and infrared mapping spectrometer
Radar altimeter
Mars observer camera
Pressure modulator infrared radiometer
Mars horizon sensor
Gamma ray spectrometer
Nadir

Figure 11-5. Configuration of instruments on the Mars Observer Spacecraft during the mapping orbit.

booms on the spacecraft (fig. 11-5). All other instruments are rigidly mounted to the spacecraft structure. No movable platform is provided; the spacecraft is continuously nadir-pointed, rotating at orbital rate. Thus, instruments that require scanned or multiple fields of view have internal scanning mechanisms.

Gamma Ray Spectrometer and Neutron Spectrometer

The team leader for the Gamma Ray Spectrometer experiment is W. Boynton of the University of Arizona. The instrument has a high spectral-resolution germanium detector cooled below 100 K by a passive radiator. It will measure the intensities of characteristic gamma-ray lines that emerge from the Martian surface within the energy range 0.20 to 10 MeV. Neutrons will be detected with a plastic scintillator to determine the abundance of H_2O and CO_2.

The specific objectives of this investigation are to

1. Determine the elemental composition of the surface of Mars with a spatial resolution of a few hundred kilometers through measurements of excited gamma-rays and albedo neutrons (H, O, Mg, Al, Si, S, Cl, K, Ca, Fe, Th, and U);

2. Determine hydrogen depth dependence in the top tens of centimeters of the surface;

3. Determine the atmospheric column density; and

4. Determine the arrival time and spectra of gamma-ray bursts.

Mars Observer Camera

The principal investigator of the Mars Observer Camera is M. Malin of Arizona State University. The instrument is a line scan camera which incorporates both wide-angle and narrow-angle optics for producing global coverage (7.5 km/pixel), selective moderate resolution images (480 m/pixel), and very selective high-resolution (1.4 m/pixel) images.

The specific objectives of this investigation are to

1. Obtain global synoptic views of the Martian atmosphere and surface to study meteorological, climatological, and related surface changes;

2. Monitor surface and atmospheric features at moderate resolution for changes at time scales of hours, days, weeks, months, and a year; and

3. Systematically examine local areas at extremely high spatial resolution in order to quantify surface/atmosphere interactions and geological processes.

Visual and Infrared Mapping Spectrometer

This instrument was selected for the Mission, but the necessity to reduce anticipated costs to meet the budget led to its deletion in August 1988.

The team leader for the Visual and Infrared Mapping Spectrometer was L. Soderblom of the U.S. Geological Survey. The instrument was a whisk broom, 320-channel mapping spectrometer operating in the 0.35 to 5.2 micrometer spectral region. It utilized a line array cooled to 80 K by a passive radiator, grating dispersion, and had a 1.8 mrad instantaneous field-of-view and a full field-of-view of 6.6°. The high spectral resolution and broad spectral coverage (11 nanometers from 0.35 to 2.4 micrometers and 22 nanometers from 2.4 to 5.2 micrometers) would have permitted direct identification of many minerals on Mars.

The specific objectives of this investigation were to

1. Produce km-resolution mosaics of the Martian surface in 320 spectral channels for the purpose of identifying mineralogical and chemical units, studying the distribution of surface volatiles, and understanding the physical structure of the regolith; and

2. Produce a global map of the Martian surface at 10 km resolution in 10 selected wavelengths to extend the local interpretation to a global scale.

Thermal Emission Spectrometer

The principal investigator of the Thermal Emission Spectrometer is P. Christensen of Arizona State University. The instrument is a Michelson interferometer which covers the spectral range 6.25 to 50 micrometers with 10 cm^{-1} spectral resolution. Separate solar reflectance (0.3 to 3.9 micrometers) and broad band radiance (0.3 to 100 micrometers) channels are included. It has six 8.3 mrad fields of view, each with 3 km spatial resolution at nadir.

The specific objectives of this investigation are to

1. Determine and map the composition of surface minerals, rocks, and ices;

2. Study the composition, particle size, and spatial and temporal distribution of atmospheric dust;

3. Locate water-ice and carbon dioxide condensation clouds and determine their temperature, height, and condensate abundance;

4. Study the growth, retreat, and total energy balance of the polar cap deposits;

5. Measure the thermophysical properties of the Martian surface (thermal inertia, albedo) that can be used to derive surface particle size and rock abundance; and

6. Determine atmospheric temperature, pressure, water vapor, and ozone profiles, and seasonal pressure variations.

Pressure Modulator Infrared Radiometer

The principal investigator of the Pressure Modulator Infrared Radiometer is D. McCleese of the Jet Propulsion Laboratory. The instrument is a limb, off-nadir, and nadir scanning radiometer. Measurements are made in nine spectral bands with five filter channels and two pressure modulator cells (one containing carbon dioxide, the other, water vapor). The detectors are cooled to 88 K by a passive radiator.

The specific objectives of this investigation are to

1. Map the three-dimensional and time-varying thermal structure of the atmosphere from the surface to 80 km altitude;

2. Map the atmospheric dust loading and its global, vertical, and temporal variation;

3. Map the seasonal and spatial variation of the vertical distribution of atmospheric water vapor to an altitude of at least 35 km;

4. Distinguish between atmospheric condensates and map their spatial and temporal variation;

5. Map the seasonal and spatial variability of atmospheric pressure; and

6. Monitor the polar radiation balance.

Radar Altimeter and Radiometer

This instrument was initially selected for the Mission. However, the necessity to reduce anticipated costs to meet the budget led to plans to substitute this instrument with a reduced capability and less costly laser altimeter in August 1988.

The principal investigator of the Radar Altimeter and Radiometer is D. Smith of the Goddard Space Flight Center. The instrument is a Ku-band (13.6 GHz) radar altimeter/radiometer with an adaptive resolution tracking system; it will return amplitude, shape, and time delay of echo along with brightness temperature using a 1-meter nadir-pointed antenna.

The specific objectives of this investigation are to

1. Provide topographic height measurements with a vertical resolution better than 0.5% of the elevation differences within the footprint;

2. Provide slope information, averaged over the footprint;

3. Provide surface brightness temperatures at 13.6 GHz with a precision of better than 2.5 K; and

4. Provide well-sampled, radar-return waveforms for precise range corrections and the characterization of the Martian surface.

Radio Science

The team leader for the Radio Science investigation is G. L. Tyler of Stanford University. The instrument is the spacecraft radio subsystem, x-band up- and down-link, supplemented with an ultrastable oscillator to maximize the science during occultation.

The specific objectives of this investigation are

Atmosphere–

1. Determine profiles of refractive index, number density, temperature, and pressure at about 200 m resolution for the lowest few scale heights at high latitudes in both hemispheres on a daily basis;

2. Monitor both short-term and seasonal variation in atmospheric stratification;

3. Characterize the thermal response of the atmosphere to dust loading;

4. Explore the thermal structure of the boundary layer at about 10 m resolution;

5. Determine the height and peak plasma density of the daytime ionosphere; and

6. Characterize the small-scale structure of the atmosphere and ionosphere.

Gravity–

1. Develop a global, high-resolution model of the gravitational field;

2. Determine both local and broad-scale density structure and stress state of the Martian crust and upper mantle;

3. Detect and measure temporal changes in low degree harmonics of the gravitational field; and

4. Provide planetary radius measurements.

Magnetometer and Electron Reflectometer

The principal investigator of the Magnetometer is M. Acuna of the Goddard Space Flight Center. The instrument has two triaxial fluxgate magnetometers and an electron-reflectometer, mounted on a 6-meter spacecraft boom. Two to sixteen vector samples per second will be acquired.

The specific objectives of this investigation are to

1. Establish the nature of the magnetic field of Mars;

2. Develop models for its representation which take into account the internal sources of magnetism and the effects of the interaction with the solar wind;

3. Map the Martian crustal remnant field using the fluxgate sensors and extend these *in situ* measurements with the remote capability of the electron-reflectometer sensor;

4. Characterize the solar wind/Mars plasma interaction; and

5. Remotely sense the Martian ionosphere.

Interdisciplinary Scientists

R. Arvidson of Washington University—Geoscience.

The objectives of this investigation are to gain an understanding of the mechanisms of weathering, their temporal variations, and the cycling of volatiles through the sedimentary system. He has additional responsibilities for science data management and archiving.

M. Carr of U.S. Geological Survey—Geoscience.

The objectives of this investigation are to gain a better understanding of the role of water in the evolution of the Martian surface, to characterize the planet's volcanic history, and to determine the nature and cause of the uplands/plains dichotomy.

A. Ingersoll of the California Institute of Technology—Polar Atmospheric Science.

The objectives of this investigation are to define atmospheric circulation during all seasons to specify polarward transport of carbon dioxide, water, dust, and energy as well as the radiative and surface fluxes in the polar regions.

B. Jakosky of the University of Colorado—Surface-Atmospheric Science.

The objectives of this investigation are to determine the nature of the interaction between the surface and atmosphere in order to better understand the processes involved in the formation and evolution of the Martian surface and atmosphere.

J. Pollack of Ames Research Center—Climatology.

The objectives of this investigation are to assess the influence of dust on atmospheric circulation, the factors which control the life cycle of dust storms, the role of dynamics in the seasonal water cycle, the transport of dust, the constraints on an early dense carbon dioxide atmosphere, and the modulations of atmospheric circulation due to astronomical variations.

Mission Operations and Data Analysis

The Mars Observer Mission will be conducted in a mission support area at the Jet Propulsion Laboratory and will be supported by the Deep Space Network and the Space Flight Operations Center. The science investigation teams will be remotely located at the home institutions of the principal investigators, team leaders, and other key science personnel. Workstations and electronic communication links will connect the mission planning and data analysis activities of these scientists, engineers, and mission managers. Via the workstations, the science teams will receive data and planning products, plan investigation activities, prepare inputs to the sequence generation process, provide analyses of instrument health and performance, and return reduced or processed investigation data for use by other project participants.

The Mars Observer project will use data standards for packet telemetry and telemetry channel coding and will use a standard formatted data unit for data transfer among ground systems. Mission data will be stored in a project data base. Analysis data will consist of a record of each instrument's packet telemetry data provided as an experiment data record; spacecraft position and pointing information data available as a supplementary experiment data record; and related data such as spacecraft status, commands, data availability, and ancillary data. Planning products available will be the up-to-date mission sequence plan, schedules and commanding opportunities, and orbit/viewing forecasts. Investigators and analysts will access the data base to participate in the planning process and provide analysis products, including spacecraft and instrument performance and status, and higher-order data products, such as intermediate and final science products.

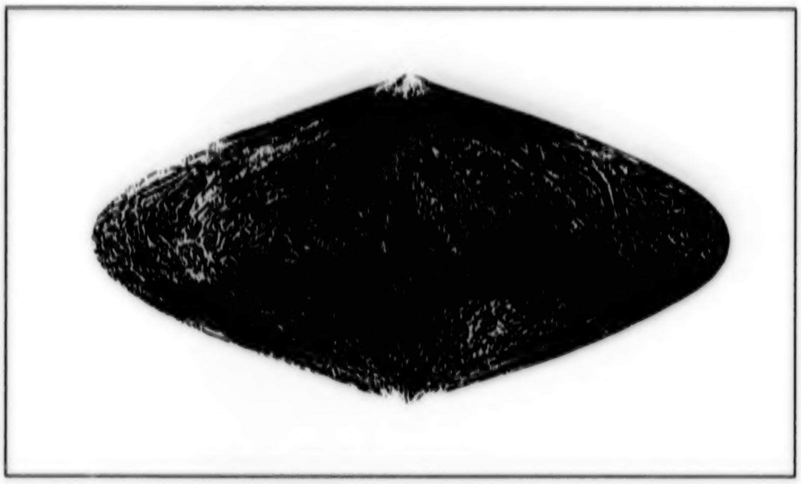


Figure 11-6. Sinusoidal equal-area projection of Mars made from a digitized airbrush map with 3.7 km pixels.

All of the instruments are "mapping" instruments in some sense. Efficient comparison of such data requires global digital data bases that are each accessible to end users and that can be manipulated without the assistance of technological experts. A sinusoidal equal-area projection will be used as a base. In this projection each parallel of latitude is an image line with its length compressed by the cosine of its latitude (fig. 11-6). A global digital image mosaic is being constructed from selected Viking images, which will be radiometrically, photometrically, and geometrically transformed. Each pixel represents 1/256° (about 230 m) and the scale can be readily changed by negative powers of two. Such a base and system is designed to allow efficient computer storage and management, user access, coregistration of data bases, rapid manipulation of data bases for effective analysis and interpretation, and inexpensive preparation of image maps on desired projections. The data bases will be widely distributed, probably on digital-optical (CD-ROM) disks.

Summary

The Mars Observer Mission will extend the exploration and characterization of Mars by providing new and systematic measurements of the surface and atmosphere of the planet. These measurements will be made from a low-altitude polar orbiter over a period of 1 Martian year, permitting repetitive observations of the surface and of the seasonal variations of the atmosphere. The Mission is being designed and will be conducted in a manner that will provide new and valuable scientific data at a significant reduction in cost and operational complexity.

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12

CRAF Mission: An Opportunity for Exobiology

N93-18558

M. Neugebauer and P. Weissman

The Halley missions of 1986 gave us a first, quick glimpse of a comet nucleus and the first *in situ* measurements of cometary gas and dust. Many of our basic ideas about cometary nuclei were confirmed while a number of startling new discoveries were also made. However, in many respects the very fast Halley flybys raised more questions than they answered. We learned, for example, that comets contain a large amount of organic material, but we were unable to determine precisely which organic molecules were present. We learned, too, that the nucleus of a comet is a dark, irregularly shaped body, but we could determine very little about the physical state and structure of the ices and grains within the comet nucleus.

On NASA's drawing boards are detailed plans for a space mission to a comet that will answer many of these questions. The Mission is called the Comet Rendezvous Asteroid Flyby, or CRAF. The term rendezvous indicates that the spacecraft will be maneuvered so that it will follow an orbit around the Sun that precisely matches the

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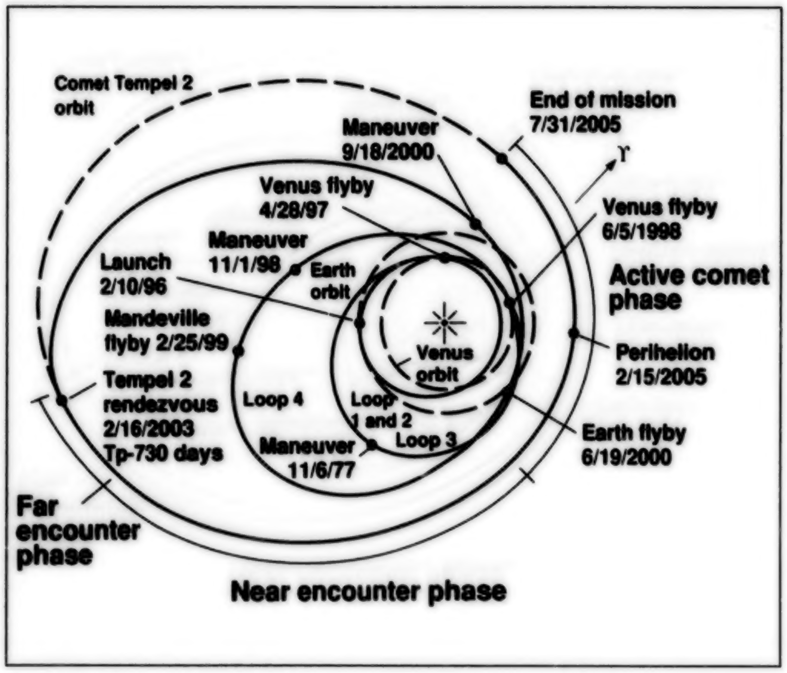
comet's orbit; the spacecraft and the comet will then travel together through one or more complete orbits. En route to the comet, CRAF will also fly by a main belt asteroid and make remote sensing measurements of its properties.

CRAF represents a substantial advance beyond the fast (70 kilometers/second) Halley flybys because a rendezvous mission has the ability to observe the comet on a very fine scale and over a very long period of time, thus allowing studies of the effects of the changing levels of the comet's activity. With a rendezvous mission it is also possible to collect and study cometary material without damaging it and to carry out a program of adaptive exploration so that early results can be used to plan later measurements.

The planned target for CRAF's rendezvous is either comet Kopff or comet Tempel 2, depending on launch date. These are two of the more active of those short-period comets which can be reached for a rendezvous mission. An active comet is preferred over one that has spent so much time in the inner solar system that it has largely exhausted its supply of volatiles. Furthermore, a high level of activity will assure that a substantial supply of gas and dust will reach the orbiting spacecraft for highly accurate *in situ* studies.

A representative CRAF trajectory is shown in figure 12-1. For a launch in February 1996, the spacecraft would initially travel on a complex trajectory that takes it twice past Venus and once past the Earth; the gravity of Venus and Earth will be used to modify the orbit and give the spacecraft sufficient energy to meet up with the target comet. The trajectory to comet Tempel 2 shown in figure 12-1 takes the spacecraft through the asteroid belt, where it will make a close flyby of the asteroid 739 Mandeville. This asteroid, whose diameter is about 110 kilometers, is a member of the primitive type CP spectral class with a surface composition believed to be similar to that of carbonaceous chondrite meteorites. All of the remote sensing instruments (described below) will operate during the asteroid flyby. CRAF will be able to determine the size, shape, mass, and density of the asteroid and obtain data concerning its surface morphology, temperature, and composition. Later launch opportunities to comet Kopff have trajectories similar to the trajectory to Tempel 2 shown in figure 12-1, but the opportunities differ in the asteroids that can be encountered.

Figure 12-1. The orbit of comet Tempel 2 and a representative trajectory of the Comet Rendezvous Asteroid Flyby (CRAF) spacecraft.


Comet Tempel 2 orbit
End of mission 7/31/2005
Maneuver 9/18/2000
♈
Venus flyby 4/28/97
Venus flyby 6/5/1998
Maneuver 11/1/98
Launch 2/10/96
Earth orbit
Active comet phase
Mandeville flyby 2/25/99
Venus orbit
Perihelion 2/15/2005
Tempel 2 rendezvous 2/16/2003 Tp-730 days
Loop 4
Loop 1 and 2
Loop 3
Maneuver 11/6/77
Earth flyby 6/19/2000
Far encounter phase
Near encounter phase

For the example mission described above, the CRAF spacecraft would finally reach its destination and be inserted into the same orbit about the Sun as comet Tempel 2 in January 2003. If CRAF is launched in May 1997, it would arrive at comet Kopff in December 2005. In either case, at the time of rendezvous, the comet will be near its aphelion and in a state of minimum activity. From a close orbit around the comet, CRAF will map the entire surface with a battery of imaging and spectroscopic instruments. Multicolor images with a resolution of better than 1 meter will provide details about the size, shape, and surface morphology of the comet nucleus. By analyzing the spectrum of reflected sunlight at visible and near-infrared wavelengths, we will be able to identify different ices and minerals on the comet's surface. The surface temperature distribution will be mapped to determine thermal properties and energy balance as well as to aid in the identification of surface materials. Accurate radio tracking data will be used to determine the mass and higher order gravity harmonics of the comet nucleus. Combined with the imaging data, knowledge of the mass will enable calculation of the comet's bulk density to tell us whether the interior of the nucleus more closely resembles a hard-packed snowball or a newly fallen bank of fluffy snow. This will then tell us how the comet nucleus probably accreted and what processing it might have undergone since its formation.

As it approaches the Sun, the comet will become more active and the spacecraft will move in and out through the comet's atmosphere, collecting dust for on-board analysis of its elemental and chemical properties. Emitted gases will be analyzed for their composition and temperature, and the velocity at which they flow away from the nucleus. The cometary atmosphere and nucleus will be studied as the comet moves along its eccentric orbit, reaching maximum activity near perihelion. During this period of high cometary activity a complete suite of plasma instruments will study the interactions of cometary gas and dust with sunlight and with the solar wind in an effort to achieve better understanding of the processes that control the ever-changing cometary tails, the acceleration of energetic particles, and other important astrophysical phenomena.

A computer drawing of the CRAF spacecraft design is shown in figure 12-2. The spacecraft consists of a central structure, or "bus," containing the electronics, radio transmitters and receivers, computers, and solid state recorders. Above the bus is a 4-meter high-gain antenna supplied by the Italian Space Agency; the antenna is kept pointed at the Earth during most of the Mission. Below is the propulsion system consisting of fuel tanks, main engines, and attitude control thrusters; the propulsion module will be supplied by the Federal Republic of Germany which will also supply one of the scientific instruments. Table 12-1 lists the scientific experiments to be carried on CRAF. Most of the instruments are located on platforms at the ends of booms extending to the right and left of the bus. On the right is the high-precision scan platform, capable of an inertial pointing accuracy of 2 milliradians (0.1 degrees),

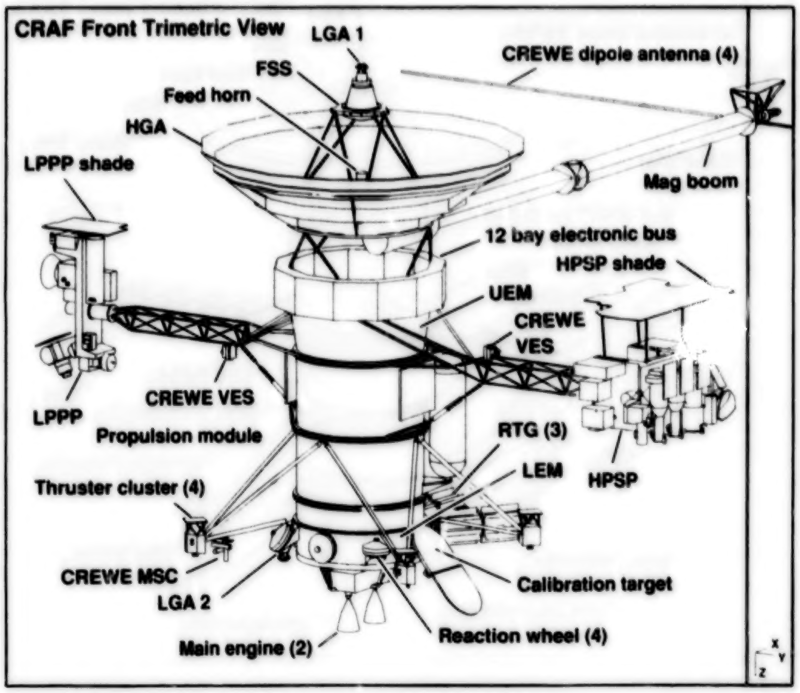

CRAF Front Trimetric View
LGA 1
FSS
Feed horn
HGA
LPPP shade
CREWE dipole antenna (4)
Mag boom
12 bay electronic bus
HPSP shade
UEM
CREWE
VES
CREWE VES
LPPP
Propulsion module
RTG (3)
LEM
HPSP
Thruster cluster (4)
CREWE MSC
LGA 2
Calibration target
Reaction wheel (4)
Main engine (2)
X
Y
Z

Figure 12-2. The CRAF spacecraft.

and carrying narrow- and wide-angle CCD cameras, a visual and near-infrared imaging spectrometer, and a thermal infrared radiometer. The low precision scan platform on the left has a pointing accuracy of 1.0 degree and carries the gas and dust sampling and analysis instruments. Magnetometers and plasma wave antennas are mounted on an extendable mast behind the bus. Three radioisotope thermoelectric generators are mounted on a short boom behind the bus and provide nearly 700 watts of power.

In addition to the science teams that provide instruments, the CRAF science team includes the five Interdisciplinary Scientists (IDS) identified in table 12-2. The role of the IDS for exobiology is to educate the instrument teams concerning the exobiological importance of their experiments, to help optimize the instrument designs and the exploration strategy for the study of exobiology, and to aid in the synthesis and interpretation of results in an exobiological context.

The foregoing discussion has highlighted the principal features of the CRAF Mission in a very general way. In the rest of this paper, we focus on and describe in greater detail those CRAF measurements which are expected to be of the greatest relevance to exobiology.

The CRAF measurements aimed at deeper understanding of chemical evolution in comets are symbolically summarized in figure 12-3. The outer arcs of the figure summarize the three principal questions to be addressed: (1) What biogenic materials are present in comets? (2) To what extent, and how, have these materials been synthesized into complex, biochemically interesting molecules? (3) Is there any evidence that substantial amounts of cometary material have been delivered to the planets in general and to the Earth in particular?

Table 12-1: CRAF Science Investigations

Acronym	Investigation	Principal Investigator/ Team Leader	Institution
ISS	Imaging (Facility)	Joseph Veverka	Cornell University
VIMS	Visual/infrared mapping spectrometer (Facility)	Thomas B. McCord	University of Hawaii
TIREX	Thermal infrared radiometer experiment	Francisco P. J. Valero	NASA Ames Research Center
CoMA	Cometary matter analyzer	Jochen Kissel	Max-Planck-Institut für Kernphysik
CIDEX	Comet ice/dust experiment	Glenn C. Carle	NASA Ames Research Center
CODEM	Comet dust environment monitor	W. Merle Alexander	Baylor University
NGIMS	Neutral gas and ion mass spectrometer	Hasso B. Niemann	NASA Goddard Space Flight Center
CRIMS	Comet retarding ion mass spectrometer	Thomas E. Moore	NASA Marshall Space Flight Center
SPICE	Suprathermal plasma investigation of cometary environments	James L. Burch	Southwest Research Institute
MAG	Magnetometer	Bruce Tsurutani	Jet Propulsion Laboratory
CREWE	Coordinated ratio, electrons, and waves experiment	Jack D. Scudder	NASA Goddard Space Flight Center
RSS	Radio science (Facility)	Donald K. Yeomans	Jet Propulsion Laboratory

To answer the first question, we need to inventory the elemental composition of the icy, organic, and nonvolatile components of cometary material. We wish to know not only the abundances of the principal biogenic elements H, C, N, O, P, and S, but also the other elements considered important to life, including Mg, Ca, Fe, Si, Na, Cl, K, Cr, Mn, and Ni, and the biochemically active trace elements such as Co, Cu, Zn, As, Se, Mo, Cd, Sn, and I.

Table 12-2: Interdisciplinary Scientists for CRAF

Topic	Name, Institution
Asteroid and inactive nucleus	David Morrison, NASA Ames
Active nucleus and dust	Armand Delsemme, University of Toledo
Exobiology	Christopher P. McKay, NASA Ames
Coma	Walter F. Huebner, Southwest Research Institute
Solar wind interaction	D. Asoka Mendis, University of California, San Diego

For the second question, we want to know not only which molecules exist in comets, but what the circumstances of their formation and preservation were. The different minerals and the different phases of ice found in cometary nuclei carry information about the pressure and temperature at the time and location of condensation. A very porous grain structure might be indicative of surface catalysis of the synthesis of complex molecules in comets. We also wish to know whether high-activity, nonequilibrium, or chemically incompatible compounds or radicals are frozen into cometary ices, and if so, at what temperatures they are released and become active. Complete answers to the second question therefore require an inventory of the molecular composition together with knowledge of the mineralogy or crystal structure of both the volatile and nonvolatile material and the fine-scale physical structure.

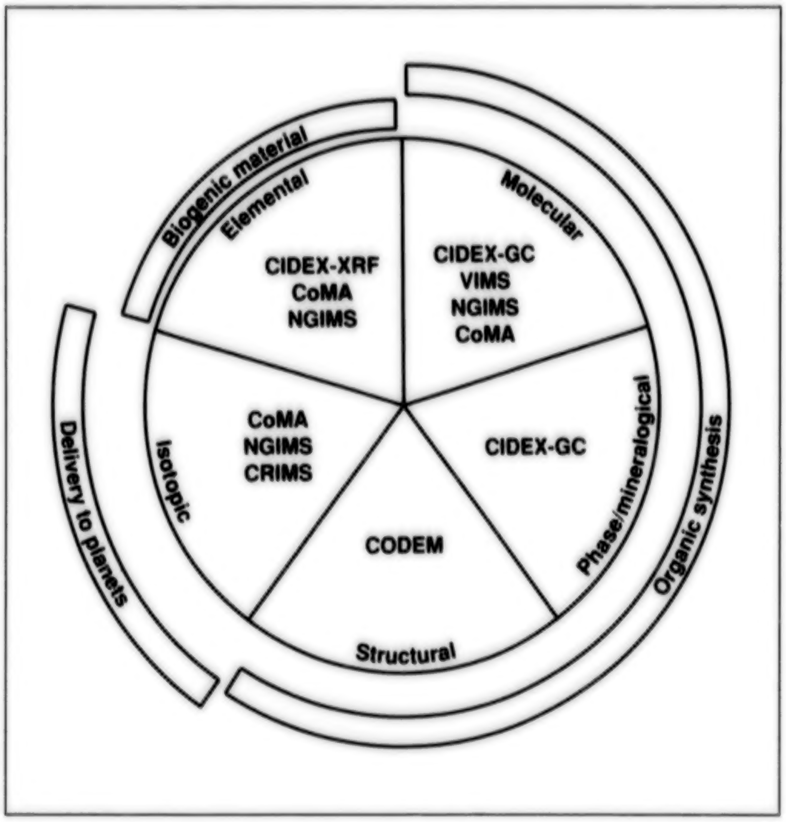


Figure 12-3. The relation between the exobiological objectives of CRAF, the parameters that need to be measured, and the CRAF instruments that make each type of measurement.

With respect to the third question, the best indicator of the contribution of cometary material to the planets probably lies in the comparison of specific isotopic ratios in comets and in the receiving medium, such as the Earth's atmosphere and oceans. The deuterium to hydrogen ratio is one of the more useful isotopic ratios for this purpose.

The pie sectors in figure 12-3 represent the five properties—elemental, molecular, and isotopic compositions together with mineralogical and physical structures—that must be measured to address the three basic questions. The acronyms and abbreviations in each of the sectors in the figure indicate which CRAF instruments contribute to measurements of each of the five properties. In the following paragraphs, we describe each of these instruments and how they will obtain the information we want.

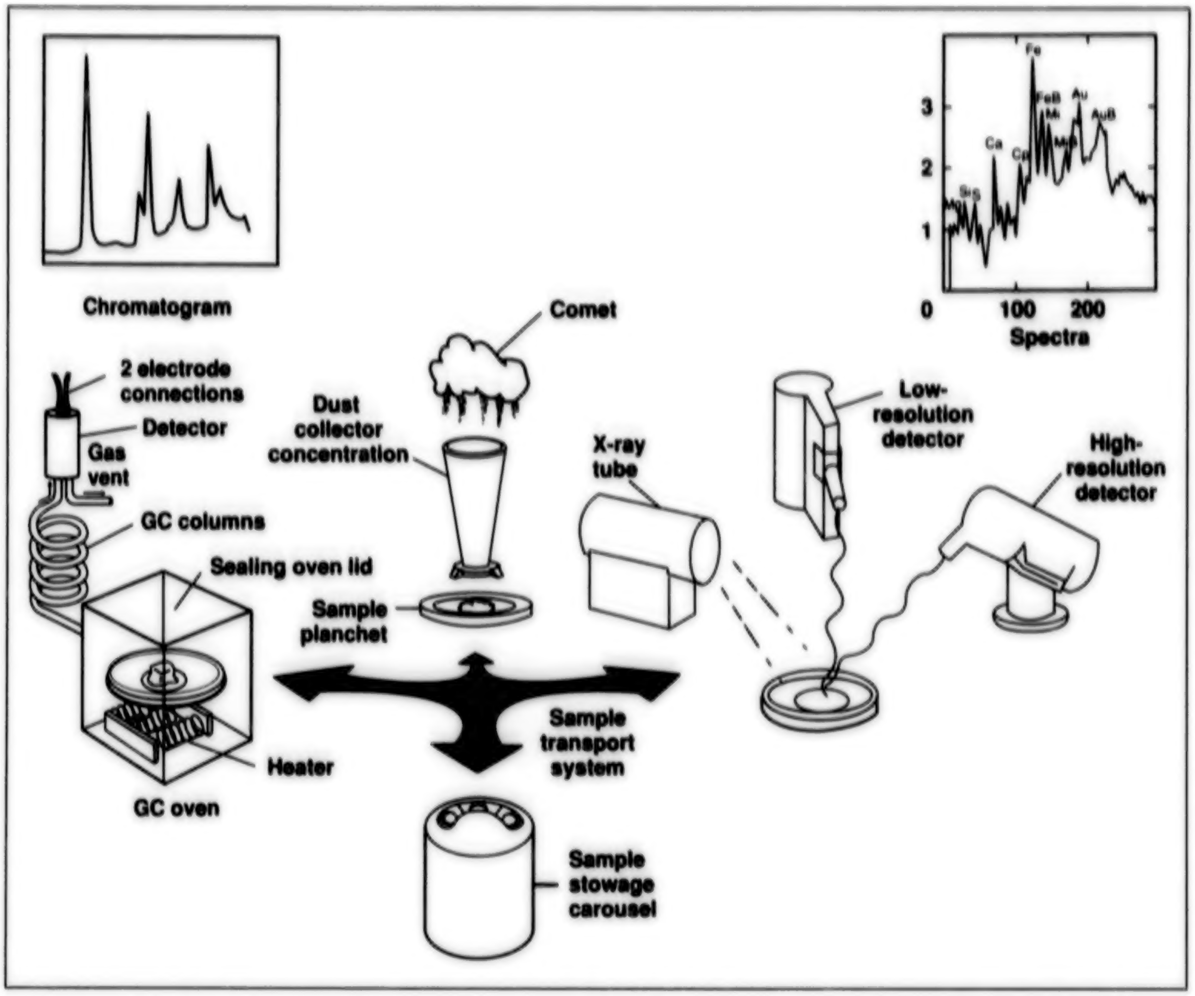

Chromatogram
Comet
0
1
2
3
100
200
Spectra
Fe
FeB
Au
Ni
AuB
NiB
Ca
Cu
Si
S
2 electrode connections
Detector
Gas vent
GC columns
Sealing oven lid
Dust collector concentration
Sample planchet
X-ray tube
Low-resolution detector
High-resolution detector
Heater
GC oven
Sample transport system
Sample stowage carousel

Figure 12-4. Diagram of the basic functions of the Cometary Ice and Dust Experiment (CIDEX).

Cometary Ice and Dust Experiment (CIDEX): The functions of CIDEX are illustrated in figure 12-4. This instrument will passively collect samples of ice and dust as the spacecraft moves in and out through the comet's coma and will then pass each sample back and forth between an x-ray fluorescence spectrometer (XRF) and a pyrolysis oven linked to a set of gas chromatograph columns (GC). In the XRF analyzer, the bulk sample will be irradiated by x-rays or alpha particles which excite secondary or fluorescent x-rays whose characteristic energies can be used to identify the abundances of 15 to 25 elements with nuclear mass of 6 or greater. After XRF analysis, a sample will be moved to a pyrolysis oven where it will be heated to a preselected temperature and the gases released during heating will be analyzed with three, possibly four, gas-chromatograph columns. There will be one column for light gases (Na, H_2, N_2, Ar, and CO), one for polar molecules (H_2O, CO_2, SO_2, H_2S, COS, HCN, CH_3CN, NH_3, CH_2O, and CH_3OH), and one

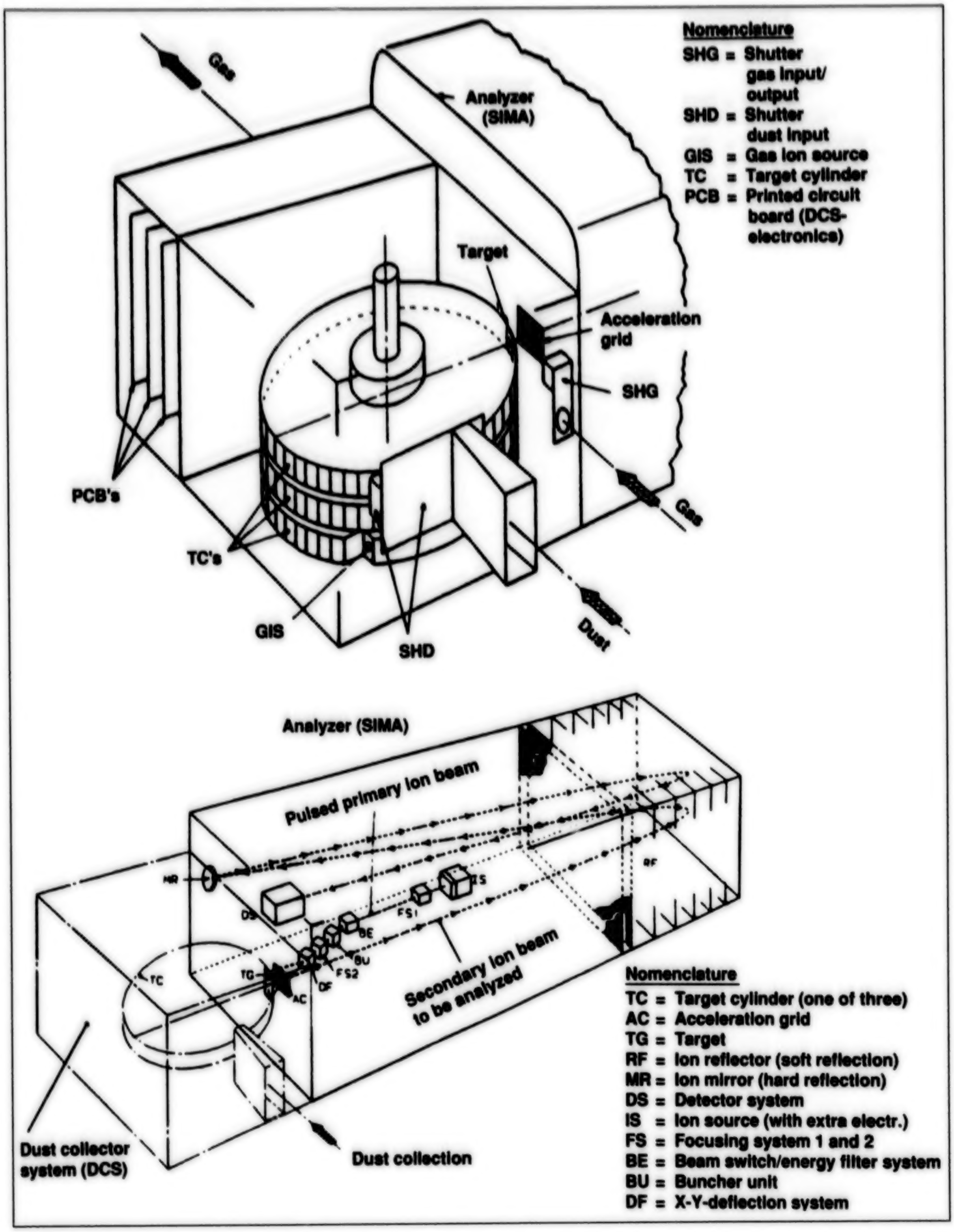
Figure 12-5. Diagram of the Visual and Infrared Mapping Spectrometer (VIMS).

for organics (all molecules with 1 to 4 carbon atoms, and some with as many as 8 carbon atoms in a molecule). The measurement cycle will consist of an XRF analysis, pyrolysis, analysis of the evolved gases with the GC, another XRF analysis to measure changes in composition, pyrolysis to a higher temperature, etc. The final step will consist of oxidizing the sample at about 800°C and then analyzing the combustion products with the GC.

Visual and Infrared Mapping Spectrometer (VIMS): VIMS will map the asteroid and the nucleus and coma of the comet in each or any of 320 spectral bands between 0.35 and 5.1 micrometers. The main elements of the VIMS instrument are shown in figure 12-5. When the comet is inactive, VIMS is expected to identify minerals and ices on the nucleus surface and to measure their abundances, distributions, and changes in time. Surface materials that can be identified by VIMS include silicates, ices, organics, oxides, salts, metals, and emitting ions. When the comet is active, VIMS will obtain maps of many of the more abundant molecules and radicals in the coma, such as CN, HCN, H_2O, H_2CO, OH, CH_4, C_2, C_3, and CO_2.

Cometary Matter Analyzer (CoMA): This instrument is based on the Secondary Ion Mass Spectrometry (SIMS) technique. A diagram is shown in figure 12-6. Samples

Figure 12-6. Diagram of the Cometary Matter Analyzer (CoMA).


Gas
Analyzer (SIMA)
Nomenclature
SHG = Shutter gas input/output
SHD = Shutter dust input
GIS = Gas ion source
TC = Target cylinder
PCB = Printed circuit board (DCS-electronics)
Target
Acceleration grid
SHG
PCB's
TC's
GIS
SHD
Dust
Gas
Analyzer (SIMA)
Pulsed primary ion beam
MR
DS
TC
TG
AC
DF
FS2
BU
BE
FS1
IS
RF
Secondary ion beam to be analyzed
Dust collector system (DCS)
Dust collection
Nomenclature
TC = Target cylinder (one of three)
AC = Acceleration grid
TG = Target
RF = Ion reflector (soft reflection)
MR = Ion mirror (hard reflection)
DS = Detector system
IS = Ion source (with extra electr.)
FS = Focusing system 1 and 2
BE = Beam switch/energy filter system
BU = Buncher unit
DF = X-Y-deflection system

of dust will be collected and then bombarded by a pulsed ion beam. The secondary ions released as a result of this bombardment will be analyzed in a time-of-flight mass spectrometer. Because of the very long path length of the mass spectrometer, very high mass resolution ($m/\Delta m \geq 3000$) can be achieved, which will allow separation of different species with the same atomic weight—such as ^{12}CH versus ^{13}C, or CO versus N_2. Thus the CoMA instrument will be very important for the measurement of isotopic ratios of cometary material. CoMA will be able to measure very heavy molecules—with molecular weights up to 3000 atomic mass units. The instrument will also be operated in different modes that enable it to measure compositional depth profiles of individual grains or to obtain mass spectra of the gas or thermal ions in the comet's coma.

Neutral Gas and Ion Mass Spectrometer (NGIMS): A diagram of NGIMS is given in figure 12-7. This instrument can analyze the gas which enters through any of four apertures:

1. An open or fly-through aperture which will use a molecular beam technique to analyze reactive species without letting them hit any walls where they might recombine;

2. A closed source, which will allow concentration of non-reactive species;

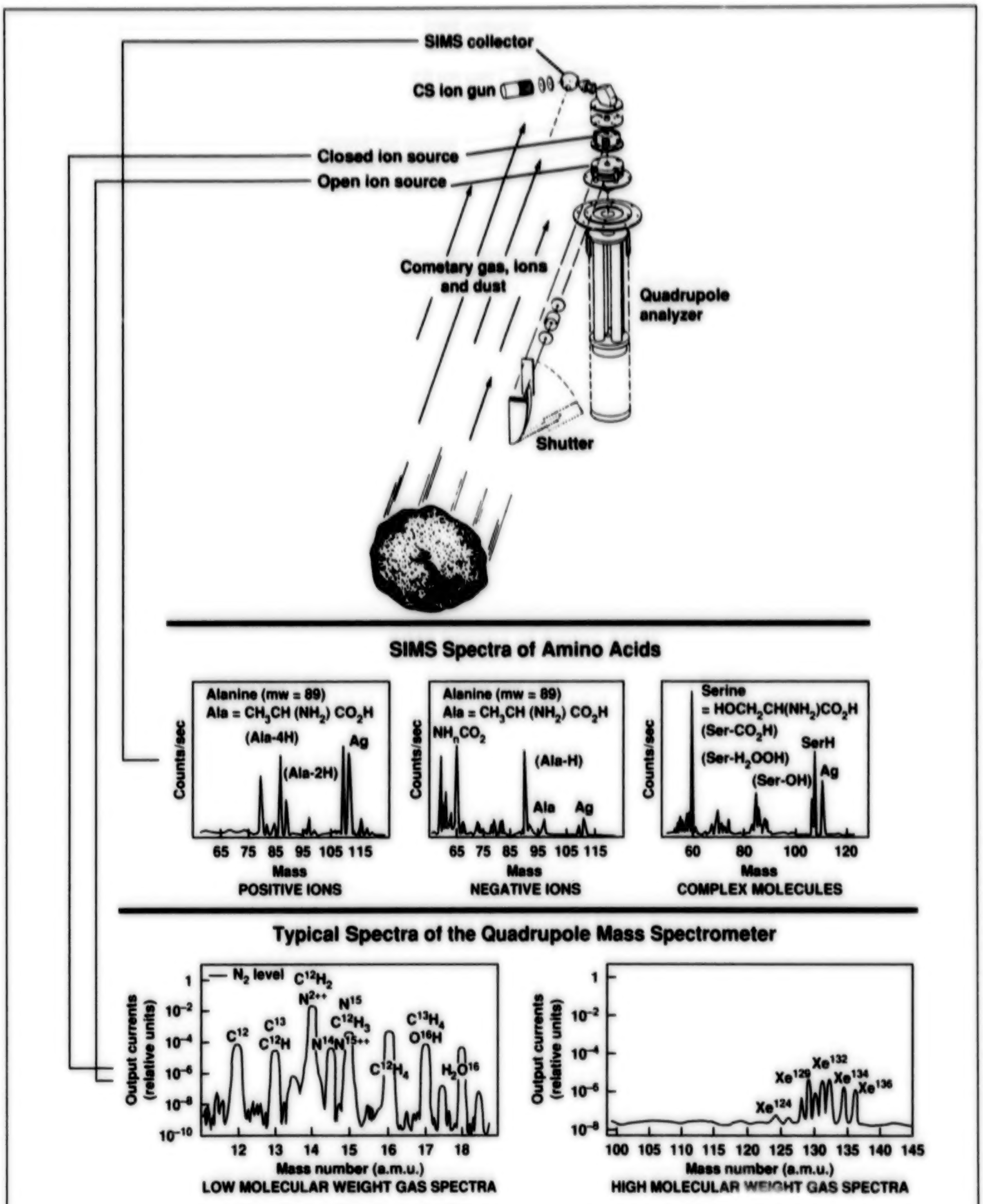
3. A thermal ion source; and

4. A SIMS source which consists of a cold metal surface that will efficiently trap heavy organic species and then occasionally gently sputter them from the surface by ion bombardment.

The mass range of this instrument is 1-300 atomic mass units. This mass spectrometer and the gas analysis mode of CoMA will obtain complementary data; the NGIMS has high sensitivity (2×10^{-3} counts/molecule/cubic centimeter) but relatively low mass resolution ($\Delta m = 0.5$ atomic mass units), while CoMA has very high mass resolution, but two orders of magnitude less sensitivity.

Cometary Retarding Ion Mass Spectrometer (CRIMS): Because most of the chemistry occurring in the comet's coma involves ion-molecule reactions, measurement of the abundance of different ion species will be of great aid in interpreting the mass spectrum obtained by NGIMS. Complete physical/chemical models of the coma will be required to interpret the observed mass spectrum in terms of chemical species. Quite sophisticated modeling will be required, for example, to sort out how much of the mass is equal to a 17 atomic mass units peak observed by NGIMS may be due to the OH radical and how much due to NH_3, or how much of the mass/charge is equal to an 18 atomic mass units/charge peak observed by CRIMS may be due to H_2O^+ and how much due to NH_4^+. Data from both NGIMS and CRIMS, as well as electron temperature data from other instruments on CRAF, are required for the unambiguous interpretation of either set of data.

Figure 12-7. Basic concept of the Neutral Gas and Ion Mass Spectrometer for the CRAF Mission.


SIMS collector
CS ion gun
Closed ion source
Open ion source
Cometary gas, ions and dust
Quadrupole analyzer
Shutter
SIMS Spectra of Amino Acids
Alanine (mw = 89)
Ala = CH3CH (NH2) CO2H
(Ala-4H)
(Ala-2H)
Ag
Counts/sec
65 75 85 95 105 115
Mass
POSITIVE IONS
Alanine (mw = 89)
Ala = CH3CH (NH2) CO2H
NHnCO2
(Ala-H)
Ala
Ag
Counts/sec
65 75 85 95 105 115
Mass
NEGATIVE IONS
Serine
= HOCH2CH(NH2)CO2H
(Ser-CO2H)
(Ser-H2OOH)
(Ser-OH)
SerH
Ag
Counts/sec
60 80 100 120
Mass
COMPLEX MOLECULES
Typical Spectra of the Quadrupole Mass Spectrometer
N2 level
C12H2
N2++
N15
C12H3
C13
C12
C12H
N14
N15++
C13H4
O16H
C12H4
H2O16
Output currents (relative units)
1
10−2
10−4
10−6
10−8
10−10
12 13 14 15 16 17 18
Mass number (a.m.u.)
LOW MOLECULAR WEIGHT GAS SPECTRA
Output currents (relative units)
1
10−2
10−4
10−6
10−8
Xe124
Xe129
Xe132
Xe134
Xe136
100 105 110 115 120 125 130 135 140 145
Mass number (a.m.u.)
HIGH MOLECULAR WEIGHT GAS SPECTRA

- To measure the chemical composition of the neutral gas in the coma of the comet.
- To measure the composition of the low-energy ions in the coma of the comet.

Cometary Dust Environment Monitor (CODEM): To determine the dust particle size distribution, CRAF will also carry an instrument to record the flux of very small solid particles. The principal detector of this instrument is a very sensitive microphone that listens for and records the magnitude and frequency of particle impacts.

Conclusion

Returning to figure 12-3, we can now comment on the apparent overlap of measurements in each of the five pie sectors. The various measurements will, in fact, be complementary, rather than redundant. For example, the several measurements of elemental composition will be complementary because the CIDEX-XRF will determine the elemental abundance of bulk samples of dust, CoMA will obtain high-resolution mass spectra of both grains and gas, while NGIMS will measure the composition of gas and thermal ions at high sensitivity, but lower resolution. Similar complementarity is found in each of the five pie sectors. Some instruments will measure the nucleus or its surface, some the gas, and some the dust. Furthermore, some types of measurements will help resolve ambiguities in other types of measurements. For example, comparison of gas chromatograph results with molecular weight spectra obtained by NGIMS and CRIMS and with observations of spectral lines of coma gases by VIMS will give us three independent types of measurements, with a much reduced chance of confusion and ambiguity.

In summary, the Comet Rendezvous Asteroid Flyby Mission provides an opportunity for significant advances in our knowledge of the primitive, yet complex material in a short-period comet and on the surface of an asteroid. CRAF will obtain new information concerning the early environments of the outer solar system and the asteroid belt and of the extent of chemical evolution in such environments. Comparison of data from the asteroid flyby and the comet rendezvous with results of studies of interplanetary dust and meteorites will elucidate the relations between these different solar system objects. The results of the CRAF experiments should also give us a clearer picture of how the primitive bodies, the comets and asteroids, and their interplanetary debris may have contributed to the origin of life on Earth.

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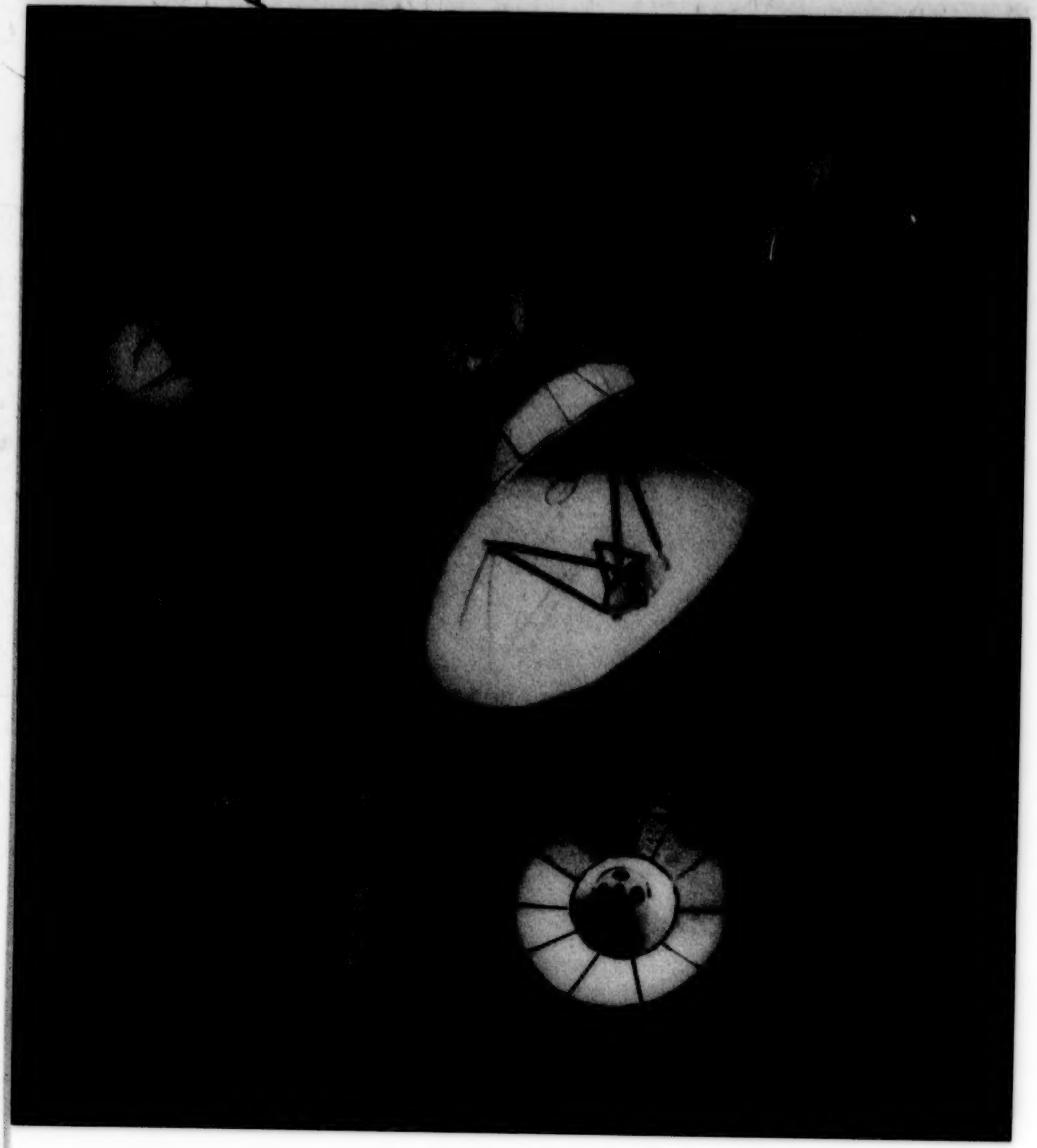

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Cassini: Mission to Saturn and Titan

S. J. Kerridge, W. Flury, L. J. Horn, J. P. LeBreton, D. S. Stetson, R. L. Stoller, and G. H. Tan

The Cassini Mission to Saturn and Titan represents an important step in the exploration of the outer planets. It will expand on the flyby encounters of Pioneer and Voyager and parallel the detailed exploration of the Jupiter system to be accomplished by the Galileo Mission. By continuing the study of the two giant planets and enabling detailed comparisons of their structure and behavior, Cassini will provide a tremendous insight into the formation and evolution of the solar system. In addition, by virtue of its focus on the Saturnian satellite Titan, Cassini will return detailed data on an environment whose atmospheric chemistry may resemble that of the primitive Earth.

The Cassini Mission, named after the Italian astronomer Domenico Cassini who first discovered structure in Saturn's rings, will consist of two major parts. The first will be the delivery of a probe into the atmosphere of

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Titan for sampling and detailed chemical and physical analysis. This will be followed by a 4-year orbital tour of all parts of the Saturn system, including numerous flybys of Titan and the "icy" satellites. NASA will build the orbiter as the second spacecraft in its Mariner Mark II series; these are modular spacecraft designed for reduced-cost exploration of the outer planets and small bodies. The European Space Agency (ESA) will provide the Titan probe, which will be integrated with the orbiter. The Mission will be launched in April 1996, on a Titan IV/Centaur expendable launch vehicle, and will arrive at Saturn in October 2002. Delivery of the probe to Titan will occur in January 2003, and the Mission will last until the end of the Saturn tour in September 2006.

Cassini has been the subject of a joint NASA/ESA Technical Assessment since 1984. A joint U.S./European science working group led by D. Gautier, W. Ip, and T. Owen has formulated detailed scientific objectives and a strawman instrument payload. At the same time, U.S. and European technical study teams have developed corresponding mission and spacecraft designs. Cassini is a candidate for a NASA New Start in FY 1990 (as part of the Mariner Mark II program, in combination with the Comet Rendezvous Asteroid Flyby Mission) while the Titan probe is a candidate for selection by ESA in November 1988. The instrument payload of the orbiter and the probe will comprise both U.S. and European experiments; science selection is planned for December 1990.

This chapter provides an overview of the Cassini Mission and its scientific objectives. Particular attention will be paid to Titan and the Titan probe phase of the Mission, to emphasize the potential contribution to the study of exobiology. The information contained in this chapter was current at the time of its preparation in August 1988. Since then the Cassini Mission was approved by Congress and has evolved significantly. However, the description contained herein remains representative of the Mission, particularly with respect to exobiology issues. Please contact the CRAF/Cassini project office at the Jet Propulsion Laboratory for further information.

Science Objectives and Payload

The model payloads of the probe and the orbiter are shown in table 13-1. The scientific objectives can be divided into five categories: Titan, Saturn, rings, icy satellites, and magnetospheres. The key area of interest to exobiologists is Titan; the other four scientific categories will be discussed briefly to provide a comprehensive overview of the Cassini Mission.

Table 13-1: Model Scientific Payload

Orbiter	Probe
Ultraviolet Spectrometer/ Imager	Descent Imager/Spectral Radiometer
Mid-Far Infrared Spectrometer	Lightning/Radio Wave Detector
Microwave Radiometer/ Spectrometer	Laser Spectrometer/Particle Size
High-Speed Photometer	Gas Chromatograph/Mass Spectrometer
Imaging Radar/Radar Sounder	Aerosol Collector/Pyrolyzer
Radio Science	Atmospheric Structure Instrument
Ion/Neutral Mass Spectrometer	Surface Science Package
Wide and Narrow Angle Cameras	Science Using Probe Engineering Subsystems
Near-Infrared Spectrometer	– Radar Altimeter
Ion Analyzer/Langmuir Probe	– Doppler Wind Experiment
Dust Analyzer	
Energetic Gas/Hot Plasma Detector	
Plasma/Radio Wave Spectrometer	
Plasma Spectrometer	
Magnetometer	

Titan

Titan was discovered in 1655 by the Dutch astronomer Christiaan Huygens, who also first postulated the existence of Saturn's rings. Titan is by far the largest Saturnian satellite; in fact, with the exception of Jupiter's moon Ganymede, it is the largest satellite in the solar system. It is also the only satellite known to possess a dense atmosphere, and it is this atmosphere that makes Titan such an important target for exploration. While its size and atmospheric density might stimulate comparisons to the terrestrial planets, at such a great distance from the Sun, Titan is much colder and richer in ices. The surface temperature is about 95 K, which is a few degrees above the melting point of methane.

It has been postulated that lakes or oceans of liquid methane or other hydrocarbons might dominate the surface, and that dense methane clouds exist in the lower atmosphere below the visible orange haze. Figure 13-1 shows a sketch of this model along with an atmospheric temperature profile. Also indicated is the altitude at which the Cassini probe will begin to sample the atmosphere. For reference, table 13-2 presents the current estimates of some of Titan's physical properties.

The presence of methane in Titan's atmosphere has been known since about 1944. It appears to be present in quantities sufficient to require continuous replenishment from methane "lakes" or subsurface degassing. The uniform orange haze obscures any atmospheric structure; it is generally agreed that this haze is composed of condensed organic compounds including polyacetylenes and cyanide polymers. About 10 such organic compounds have been identified, all of which can be derived by photochemistry of a nitrogen-methane mixture.

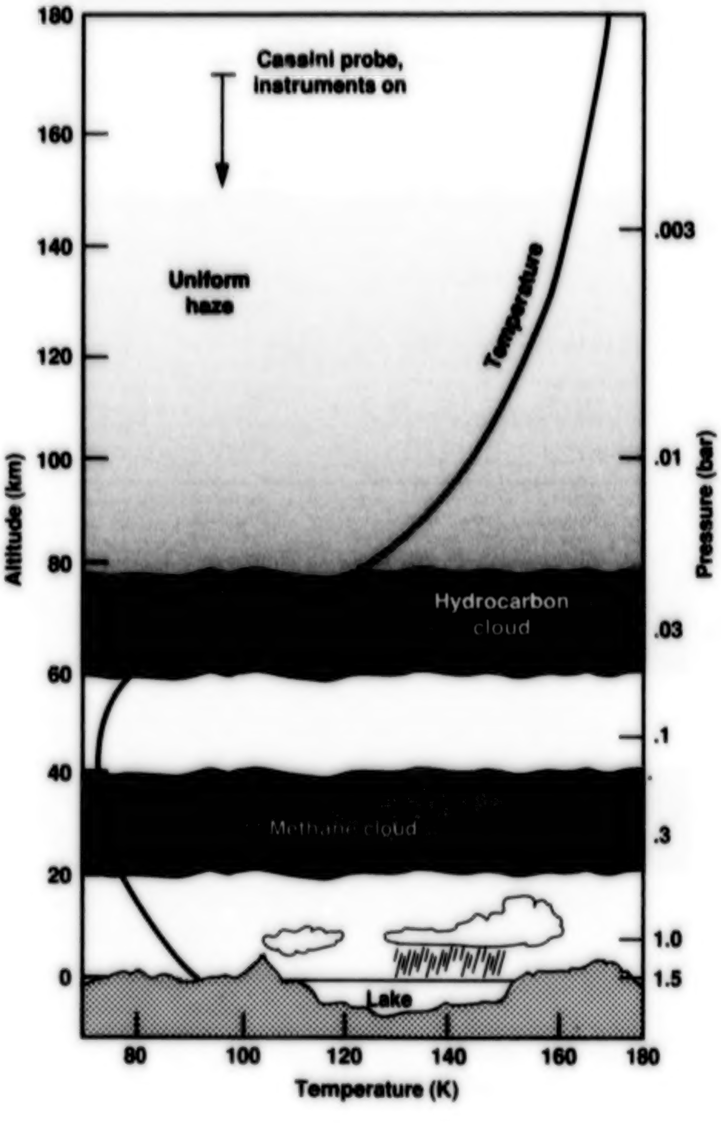
Much of the detailed information about Titan's atmosphere has been extracted from Voyager flyby data. One of the most important results was the detection of three nitriles, HCN, HC_3N, and C_2N_2. Laboratory experiments have shown that HCN is a precursor of purines (adenine in particular), which are among the building blocks of nucleic acids on Earth. Similarly, HC_3N leads to pyrimidines, which are also present in nucleic acids. Thus, some of the key outstanding issues are the abundances and distributions of gaseous nitriles in Titan's upper atmosphere, the degree of complexity these compounds have achieved, and the processes and pathways for producing these components.

Titan's stratosphere is known to contain hydrocarbons, nitriles,and oxygen-bearing compounds, while the troposphere contains CO, Ar, N_2, H_2, and CH_4. The low tropopause temperature of about 70 K, as shown in figure 13-1,

Table 13-2: Titan Physical Characteristics

Radius	2575 km
Mass	1.83 lunar masses
Rotational period	16 days
Orbital radius	20 Saturn radii
Orbital period	16 days

Figure 13-1. Titan's lower atmosphere.

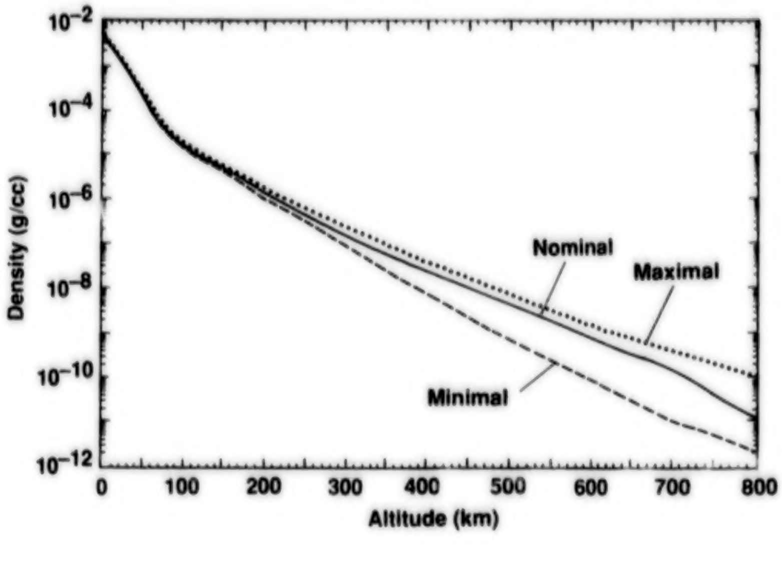

180
160
140
120
100
80
60
40
20
0
Altitude (km)
Cassini probe, instruments on
Uniform haze
Temperature
Hydrocarbon cloud
Methane cloud
Lake
.003
.01
.03
.1
.3
1.0
1.5
Pressure (bar)
80
100
120
140
160
180
Temperature (K)

acts as a cold trap for most gases and should limit their amounts in the stratosphere unless they are formed there. Thus, it appears that the stratospheric compounds are formed at this level by complex photochemistry. Condensation of these and other gases forms the thick smog-like haze and induces a greenhouse effect that is responsible for Titan's somewhat elevated surface temperature. Of critical importance to an overall understanding of Titan's atmospheric chemistry is the determination of the relative abundances of the various constituents as a function of altitude. This will serve to test the theories of atmospheric origin, and will help to determine what condensed materials should have been deposited on the surface.

While figure 13-1 shows a temperature profile in the lower atmosphere, the structure of this profile between 200 and 1500 km altitude is unknown. In addition, it has been predicted that strong zonal winds may be present, but their magnitude and direction are uncertain. These are fundamental questions which will be answered by the Titan Probe Mission; however, design of the Mission itself requires that these quantities be modeled as accurately as possible. The Cassini Mission has adopted a particular model of Titan's atmosphere, known as the Hunten-Lellouch model, for the purposes of mission and science planning. One quantity of importance to the probe design, for example, is the atmospheric density as a function of altitude, since this will determine the probe's rate of descent. Figure 13-2 shows the density predicted by the Hunten-Lellouch model.

The Cassini probe will begin sampling the atmosphere and transmitting data at an altitude of about 170 kilometers. The descent from this level to the surface will last approximately 2.5 hours. The Gas Chromatograph will measure the vertical distribution of hydrocarbons and nitriles and detect organics in gaseous form, and the Mass Spectrometer will measure elemental and isotopic abundances as a function of altitude. The Aerosol Collector and Pyrolyzer will detect condensed organics.

Figure 13-2. Hunten-Lellouch model atmosphere.



The Probe Infrared Laser Spectrometer will use a set of tunable diode lasers at selected mid-infrared wavelengths to measure volume mixing ratios for many organic compounds; it will also have a nephelometer capability to discern particle size distribution in the aerosol and cloud layers. A Descent Imager will reveal any cloud structure and atmospheric circulation. In addition, the doppler shift in the probe radio transmissions will be relayed to Earth and carefully analyzed to provide information on wind direction and magnitude during the descent.

Because of the thick haze layer, the surface of Titan has never been observed; its precise nature will remain unclear until the Cassini probe is able to resolve it. Although it is not a probe design requirement, it is expected that the probe will survive the surface impact. Provided that its radio antenna remains pointed in the general direction of the overflying Mariner Mark II spacecraft, the probe should be able to transmit several minutes of data from the surface. An accelerometer will measure the force of impact to provide some insight into the surface state, and this data will be transmitted immediately. A simple surface sample collector, such as a heated inlet, will allow a fast mass spectroscopic analysis of the surface material. Such measurements can be completed and transmitted to the orbiter in approximately 2 minutes; the radio relay link will be designed so that the orbiter will be in view of the probe for at least that long after the nominal impact time. A dedicated surface science package on the probe may consist of a refractometer to measure the index of refraction of a possible liquid surface, an X-ray fluorescence spectrometer to determine elemental composition of a solid surface, and an acoustic sounder to determine the density of the surface material.

During the 4-year Saturn orbital tour, the Mariner Mark II spacecraft will make about 40 close flybys of Titan. Many of the flyby altitudes will be below about 3000 km, and these are most useful for the radar instrument. This will allow the spacecraft to "see" Titan's surface through the orange haze; most of the surface will be mapped in this way, to establish its state (solid or liquid), topography, and structure (through radar sounding). Infrared spectra will provide vertical distributions of nitriles and other hydrocarbons on a global scale, to supplement the probe measurements.

Aeronomy instruments will sense the upper atmosphere during the closer flybys, which may be as low as 800 km altitude. The Ion/Neutral Mass Spectrometer will provide compositional information, and the Langmuir Probe will measure ion and electron temperatures and densities. Radio, solar, and stellar occultations will be used extensively to measure atmospheric abundances and physical structure.

The combination of a probe descent and multiple flybys will yield a tremendous amount of new information on the structure and behavior of Titan. The data will be analyzed by scientists from a wide variety of disciplines, including exobiology, since Titan should be an excellent natural laboratory for studying chemical processes that may have been important on the pre-biotic Earth.

Saturn

While Saturn has been observed from Earth for centuries, the state of knowledge about the planet was greatly enhanced by the Pioneer and Voyager flybys. For example, Saturn has been found to have complex atmospheric chemistry and wind patterns, an unexpectedly strong internal heat source, and a hydrogen-to-helium ratio that is significantly higher than that of Jupiter. Models have been advanced to explain these and other phenomena; validation and enhancement of these models will be made possible through detailed, long-term study by the Cassini spacecraft. A comparison of species abundance and isotopic ratio data with similar measurements by the Galileo Mission at Jupiter will advance our understanding of the differences between the two planets. This will allow more precise modeling of the composition of the pre-solar nebula, which is vital to an understanding of solar system formation and evolution.

Throughout the orbital tour, Saturn will be observed with remote-sensing instruments from a wide variety of orbits and lighting conditions. The gravitational field will be accurately mapped, and together with radio occultation data on the planet's shape, this will enable accurate modeling of the interior structure. Occultations and infrared observations will determine atmospheric structure and abundances. Determination of the global energy balance and ammonia abundance are major goals. Long-term observations of the atmosphere will permit precise modeling of circulation patterns, which is vital to an understanding of Saturn's structure.

Rings

Saturn's ring system is one of four in the outer solar system, and it is by far the most extensive and complex. The rings are composed of particles ranging in size from tiny pebbles to large boulders, including a sprinkling of fine dust. The ring particles are composed mostly of icy material, but contain impurities of unknown origin. The amount of these impurities seems to vary between the different rings.

Voyager revealed a ring system of unexpected structural complexity and variety. Images provided structural detail throughout the rings on scales as small as a few kilometers. Detailed structure, with 100-meter resolution, was visible in stellar and radio occultation data. The ring structure is driven in part by spiral density waves and spiral bending waves, and by the dynamical interactions of small moonlets with the ring particles. Other interesting structures include clumpy or "braided" rings as well as "spokes" that appear to be electrostatically charged particles levitated above the ring plane.

The Cassini orbiter will provide extensive imaging and occultation coverage of the ring system, so that the spatial and temporal variations in structure can be observed. A search for additional moonlets will be important to an understanding of ring dynamics. The composition of individual particles will be determined by reflectance spectrometry and thermal radiometry. Such a complete understanding of the planetary ring structure is required before the fundamental questions of ring age and formation can be answered.

Icy Satellites

There are eight major Saturnian satellites other than Titan; several of the larger satellites, in fact, were discovered by Domenico Cassini himself. In addition, there are several other smaller satellites and an undetermined number of moonlets near the main rings. As a class, these are termed the "icy" satellites, since they are composed primarily of water ice with varying amounts of a darker material probably similar to that found in primitive meteorites. Voyager and Pioneer showed these satellites to be remarkably diverse and to exhibit evolutionary histories unexpected for their small sizes. There are suggestions that the system has been through periods of collision and fragmentation, and that the smooth surfaces of some of the bodies imply a mechanism for internal heating and resurfacing. Indications of recent activity were present in the Voyager data and possibly corroborate the theory that the small particles of Saturn's diffuse E-ring are actually expelled by the satellite Enceladus through some unknown mechanism. Enceladus is very bright (in fact, it is the most reflective body in the Solar System), which is consistent with a surface of geologically young material.

Iapetus has a coating of dark material on its leading hemisphere which is suggestive of carbon-rich organic matter. The source of this material is unknown, and further analysis may help to elucidate the relationship among icy satellites, comets, and primitive carbonaceous meteorites. This may have important implications for the theories that invoke comets and meteorites for the initial delivery of organic molecules to the inner solar system. Exploration of Iapetus is thus given a high priority in the Cassini Mission; two very close flybys of this satellite will be included as a part of the orbital tour.

Each of the main icy satellites will be encountered several times during the Cassini Mission. Most of the flybys will be at tens of thousands of kilometers, but closer flybys of Enceladus and Dione, as well as Iapetus, are planned. Imaging, multispectral studies, and other remote sensing experiments will be active during each flyby. This will allow characterization of the surface topographies and compositions, so that the geological history of each satellite can be inferred.

Magnetospheres

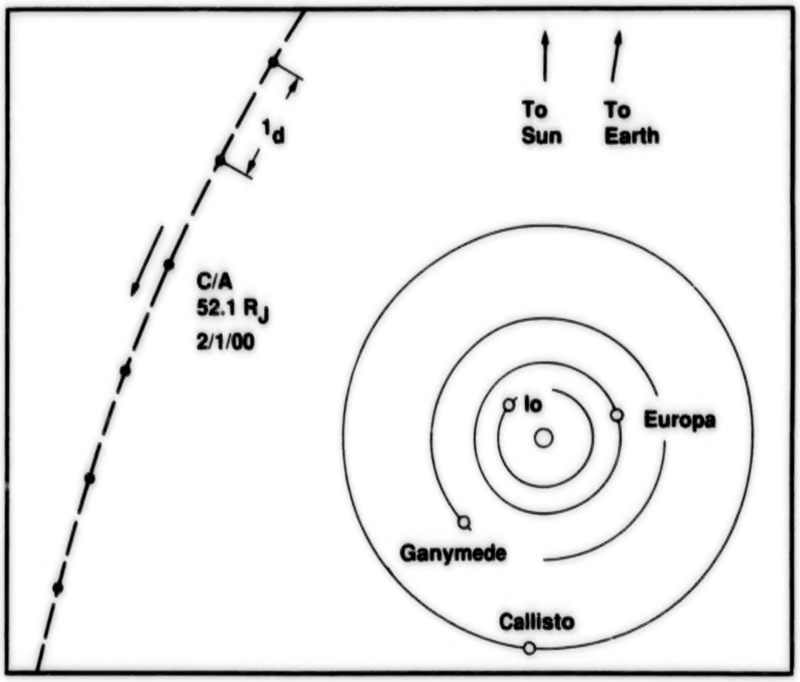
Saturn, unlike Earth, has a magnetic dipole axis that is closely aligned with the planet spin axis. The axial symmetry of the internal magnetic field is unique among the planetary magnetic fields yet observed. However, strong periodic modulations of Saturn's radio emission imply that the surface fields are not axially symmetric. The plasma environment is not well understood, though it is known to be extremely complex and time-variable, and a strong interaction between the magnetosphere and the icy satellites has been observed. Because the dimensions of the magnetosphere vary with the strength of the solar wind, Titan is at times outside the magnetosphere and in the solar wind. Titan's solar wind interactions are thought to be intermediate between those of Venus and the comets.

To thoroughly characterize the magnetosphere, the Cassini orbits will cover a wide range of orbital inclinations. Near-polar orbits will provide a direct measurement of the energy deposition into the atmosphere and the subsequent auroral emissions, and will allow study of the so-called Saturn Kilometric Radiation. Long-term observations will address the time variability of the phenomena. The Cassini plasma instruments will be on a turntable rotating at a constant 1 rpm, to provide the measurements with spatial detail. Cassini's capability for spatial and temporal resolution of the velocities of various components will place it at the forefront of planetary plasma physics.

Mission Design

The baseline Cassini interplanetary trajectory is illustrated in figure 13-3. Launch will be on April 8, 1996, using the Titan IV/Centaur launch vehicle. The fully fueled spacecraft mass will exceed the injection capability of the Titan IV/Centaur for direct trajectories to Saturn, so the

Figure 13-3. Cassini interplanetary trajectory.



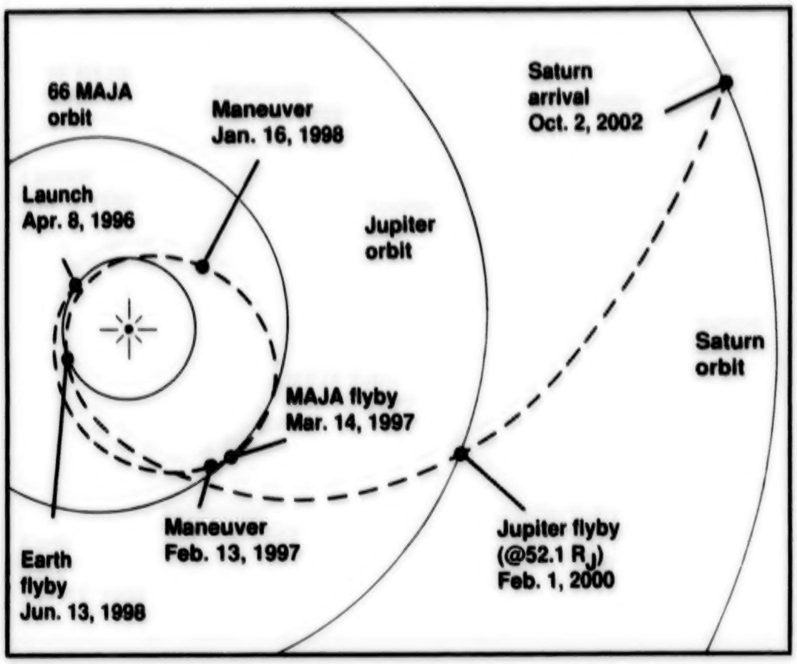

66 MAJA orbit
Maneuver Jan. 16, 1998
Saturn arrival Oct. 2, 2002
Launch Apr. 8, 1996
Jupiter orbit
Saturn orbit
MAJA flyby Mar. 14, 1997
Maneuver Feb. 13, 1997
Earth flyby Jun. 13, 1998
Jupiter flyby (@52.1 R_J) Feb. 1, 2000

Figure 13-4. Cassini Jupiter flyby.

Cassini Mission must rely on a gravity-assist flyby of Earth about 2 years after launch, and an additional flyby of Jupiter about 20 months later. The proper Jupiter-Saturn phasing will be available for missions launching in 1995-1997 and will not recur until 2016.

With the required performance margins, the current estimates of launch vehicle capability and spacecraft mass result in a minimum flight time to Saturn of about 6.5 years. This includes an allowance for a flyby of the asteroid 66 Maja, as shown in figure 13-3. Saturn arrival is currently planned for October 2, 2002.

Jupiter Flyby

Jupiter's closest approach occurs on February 1, 2000, at 52.1 Jupiter radii, a distance at which the radiation environment will pose no threat to the Mariner Mark II spacecraft subsystems. Figure 13-4 shows the spacecraft trajectory and the Galilean satellites at the time of closest approach. The flyby radius is determined by the Jupiter-Saturn phasing and the trajectory's gravity assist requirements; significantly closer approaches are not possible for the 1996 mission opportunity, so any close encounters with the Galilean satellites will be impossible. Nonetheless, significant Jupiter science can be accomplished during the flyby. For example, figure 13-5 shows the passage of the Cassini trajectory through the Jovian magnetotail. As the plot indicates, the Cassini spacecraft will spend nearly 130 days in this region. The relatively brief passages of Voyager 1 and Voyager 2 are shown for comparison. The flyby will thus allow mapping of the structure of the magnetotail in a previously unexplored region.

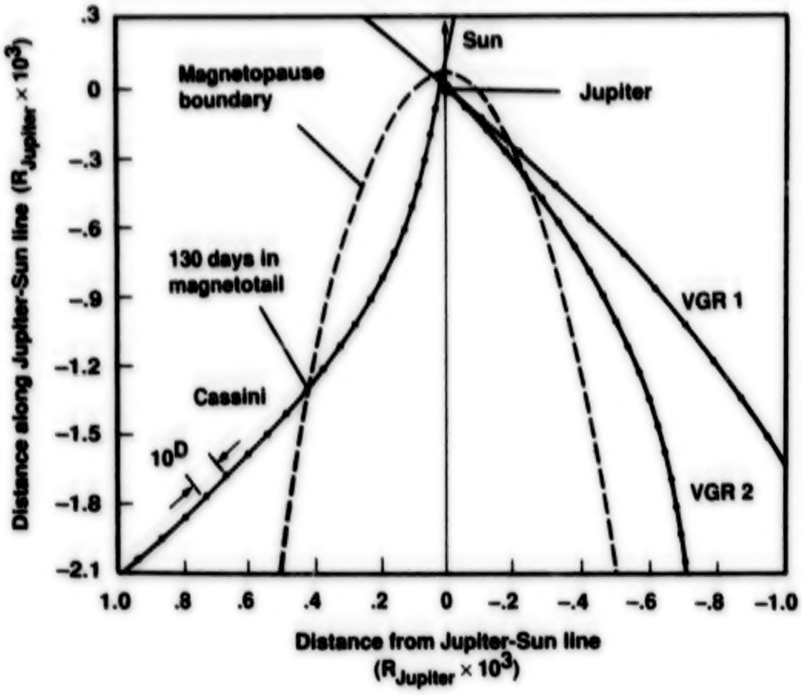


Figure 13-5. Jupiter magnetotail passages.

Saturn Orbit Insertion and Probe Delivery

Upon arrival at Saturn a "dual burn" orbit insertion strategy will be used to achieve the desired initial Saturn orbit. This strategy consists of an orbit insertion burn at periapsis (1.8 Saturn radii), which places the spacecraft into a 100-day Saturn orbit, and a periapse raise maneuver at apoapsis 50 days later. This will adjust the orbit to achieve the desired atmospheric entry conditions for the Titan probe. The probe will be released from the orbiter about 12 days before entry; the orbiter will then perform a maneuver to enable it to support the relay of the Titan probe data. To facilitate the probe design studies, ESA and NASA have agreed to limit the probe entry velocity to the range 5.9-7.1 km/sec.

Probe Entry and Data Relay

Several scientific and engineering requirements will constrain the selection of the probe's entry site. These are

1. Entry and descent on the day side of Titan;

2. Descent not more than 60° from the equator;

3. Entry favorable for doppler wind determination; and

4. Flightpath angle (at atmosphere interface) between –60 and –90°.

Figure 13-6 is a view of Titan as seen from the probe's approach trajectory, showing the allowed probe entry region.

The Saturn orbiter will be targeted to pass over (or near) the Titan probe entry point about 3 hours after probe entry. This time delay must be carefully chosen to ensure that the orbiter is properly positioned to receive and relay the probe data, including at least 2 minutes of data after the expected time of probe impact on Titan's surface. The total amount of data transmitted depends strongly on both the orbiter delay and the flyby distance. Thus the probe data relay

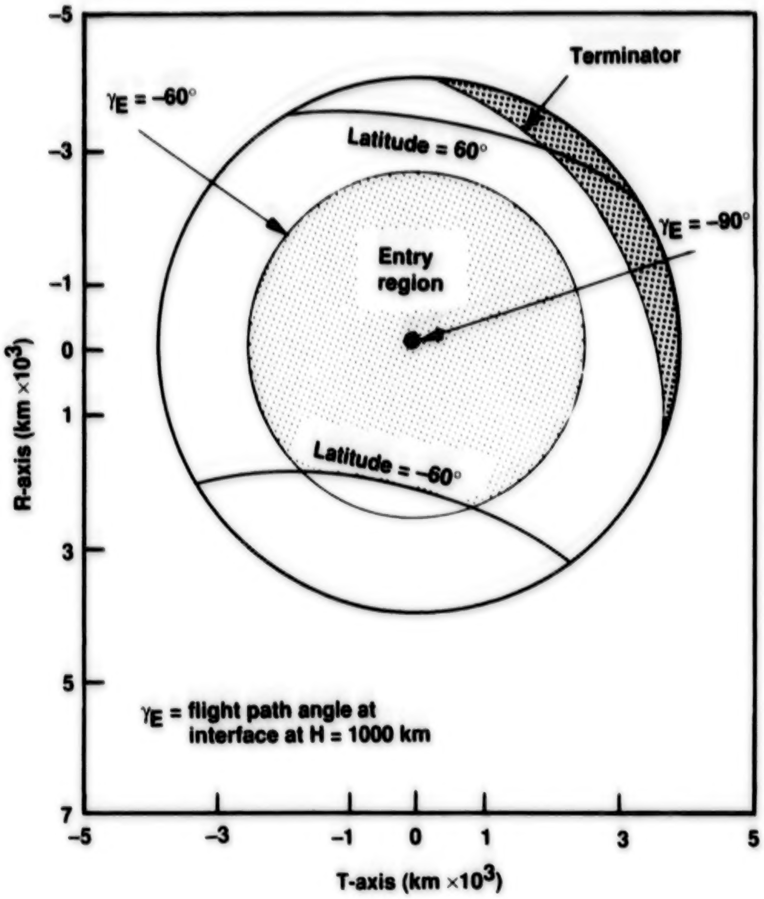


Figure 13-6. Titan probe targeting.

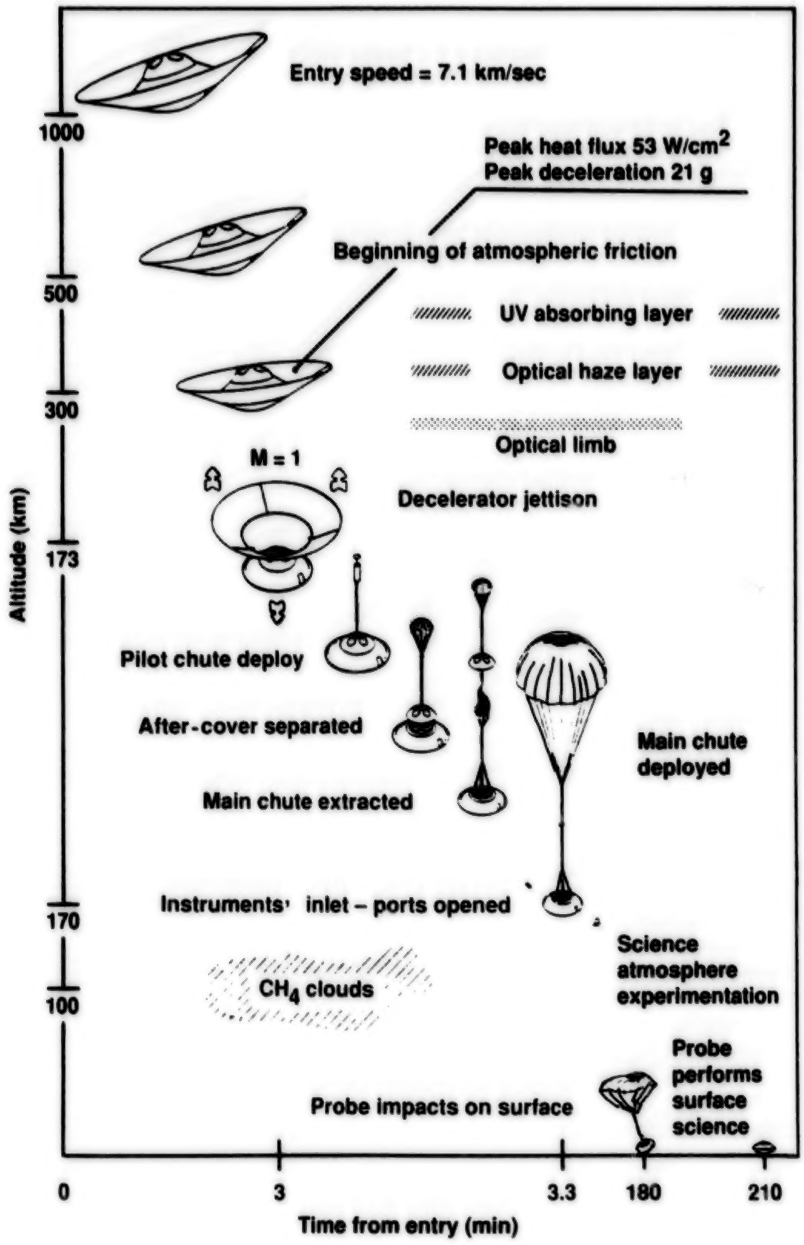
requirements constrain the flyby geometry during this initial Titan flyby. The flyby geometry must also serve to initiate the orbital tour, though, so some compromises may be required to optimize the science return from the Mission as a whole.

Figure 13-7 depicts the sequence of events that will take place during probe entry and descent. This sequence has been tailored to the Hunten-Lellouch model of the Titan atmosphere, which was previously described. The atmospheric entry interface is assumed to occur at an altitude of 1000 km; the probe will be designed to withstand an entry velocity of not more than 7.1 km/sec. The probe will descend and decelerate rapidly, reaching peak deceleration at an altitude of 230-290 km. About 3 minutes after entry, the probe will reach Mach 1.5 and the large probe decelerator will be jettisoned, marking the end of the entry phase. The only instrument active during this phase will be the Atmospheric Structure Instrument; science and engineering data will be stored within the probe for later transmission to the orbiter.

The descent phase officially begins with deployment of the first parachute. This will occur at an altitude not lower than 170 km to ensure that the presumed region of methane clouds will be sampled. All science instruments will be active as soon as the main parachute is deployed and the inlet ports are opened (see fig. 13-7). The nominal probe descent time is 164 ± 10 minutes, from parachute deployment until surface impact. While Titan's zonal winds may displace the impact point by about 300 km, they will not appreciably affect the actual descent time. The impact velocity will be about 4-7 m/sec, which the probe should easily survive. If the post-impact attitude is favorable, transmission of data from the surface will be possible; in fact, the orbiter overflight will be designed to allow at least several minutes of post-impact data relay.

During atmospheric descent, science and engineering data are transmitted to the orbiter, which transmits the data in real time and also stores them for later replay. The probe antenna beamwidth of 120° has been chosen to account for uncertainties in the probe's location and attitude during the data relay. Assuming that the transmitted bit rate can increase in discrete steps as the probe-orbiter separation decreases, the total transmitted data should exceed 23 megabits if the orbiter flyby altitude is less than 2500 km. This takes into account uncertainties in the Titan environment models and the probe and orbiter trajectories, and assumes an optimized orbited overflight time. This also allows for about 5 minutes of post-impact transmission. Preliminary relay link design studies have identified the orbiter overflight time as a critical parameter; it is strongly coupled to the duration of the atmospheric descent including uncertainties. Therefore, it is important that Titan's atmosphere and surface be modeled as accurately as possible, so that the Probe Mission's scientific return can be maximized.

Figure 13-7. Titan probe descent scenario.


Entry speed = 7.1 km/sec
Peak heat flux 53 W/cm^2
Peak deceleration 21 g
Beginning of atmospheric friction
UV absorbing layer
Optical haze layer
Optical limb
M = 1
Decelerator jettison
Pilot chute deploy
After-cover separated
Main chute extracted
Main chute deployed
Instruments' inlet – ports opened
CH$_4$ clouds
Science atmosphere experimentation
Probe impacts on surface
Probe performs surface science
Altitude (km)
1000
500
300
173
170
100
0
3
3.3
180
210
Time from entry (min)

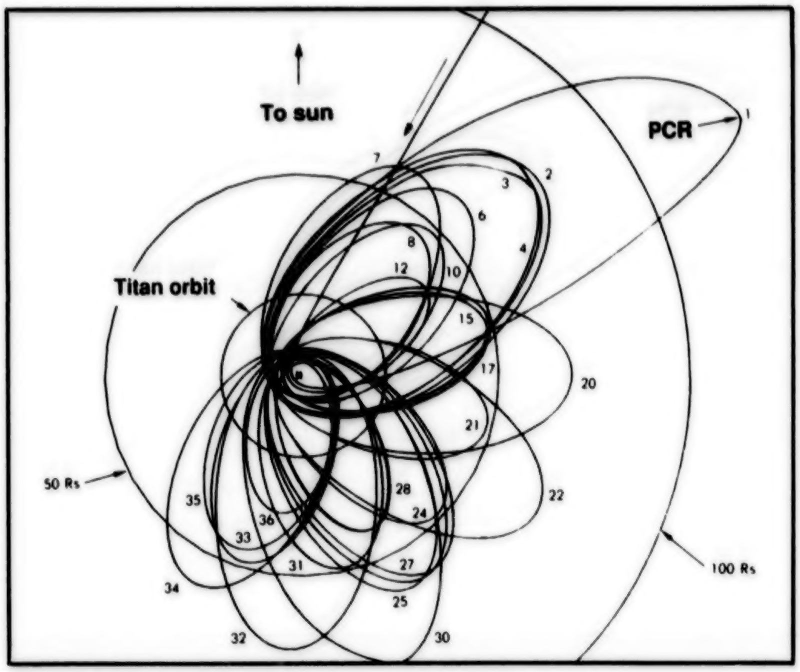


Figure 13-8. Cassini orbital tour: selected orbits.

Orbital Tour

The orbital tour phase of the Cassini Mission begins immediately after the relay of the Titan probe data and lasts until 4 years after the date of Saturn orbit insertion. The tour consists of a sequence of Saturn orbits connected by Titan gravity-assist flybys or small propulsive maneuvers; the size and orientation of these orbits is dictated by the various science requirements. Since Titan is the only Saturnian satellite massive enough to provide any appreciable gravity assist, almost every orbit must contain a Titan flyby. The tour of the Saturnian system will consist of about 40 Saturn orbits and Titan flybys; thus, there will be a great many opportunities for study of these two bodies, the "icy" satellites, the rings, the magnetosphere, and their mutual interactions.

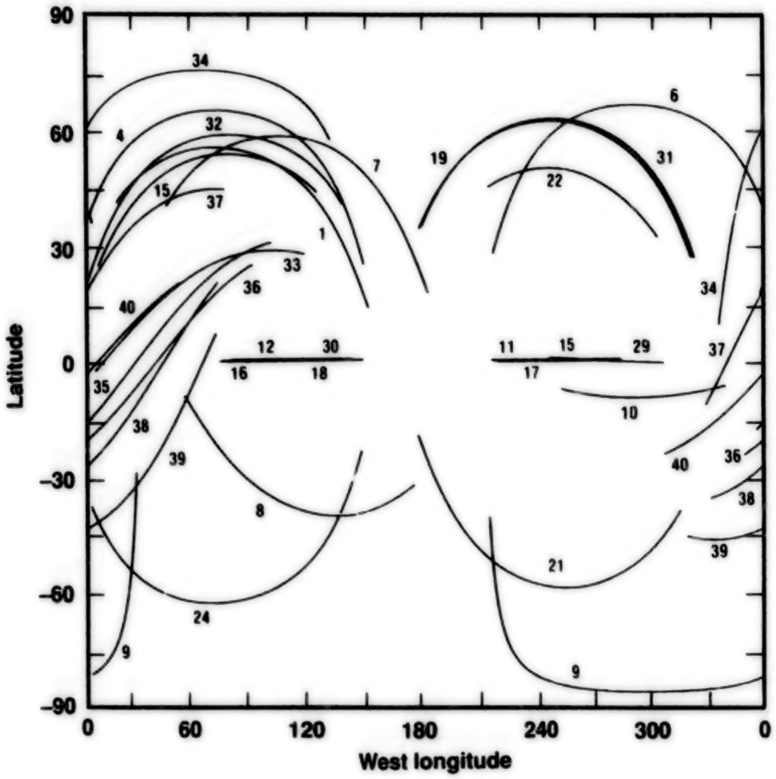
Titan gravity assists are so effective in modifying the trajectory of a Saturn orbiter that many fundamentally different types of Saturn tours are possible; there is a great deal of freedom to choose the size, inclination, and orientation of each orbit to maximize the science return. While not every possible type of tour has been studied, there is a general tour strategy which appears to offer the most satisfactory compromise among the competing scientific objectives. A tour of this type has been adopted as the baseline for the purposes of mission assessment.

Figure 13-8 shows the sequence of Saturn orbits comprising the baseline Cassini tour, with respect to a fixed Sun-Saturn line. (Note that some orbits have been omitted for clarity.) Because the figure resembles a flower, each individual orbit is referred to as a "petal." The tour begins with the initial Titan flyby, immediately upon completion of the Probe Mission. The first two or three flybys are used to reduce the orbiter's period and inclination from the long-period insertion orbit to a low-inclination orbit with a period of about 30 days. Several flybys are then used to send the spacecraft behind Saturn as seen from Earth, so that radio signals and sunlight can be used to probe Saturn's atmosphere and rings. Following these occultations, a sequence of Titan flybys removes the orbit inclination and rotates the orbit petal in the clockwise direction around Saturn. During this period, close flybys will likely be performed with Iapetus, Enceladus, and Dione.

As the orbit petal is rotated toward the anti-Sun direction, a second series of low-inclination orbits allows additional Saturn occultations, ring imaging, and a second close Iapetus flyby. The orbit is then rotated to establish the proper orientation for high-inclination Saturn occultations, and the last 10 or 12 Titan flybys are used to raise the inclination up to a final value of at least 80°. The orbital mechanics of the tour require that, as the inclination is raised, the orbital period must be reduced. The tour ends 4 years after Saturn orbit insertion, with the spacecraft in a high-inclination orbit with a period of 7.1 days.

Most of the approximately 40 Titan Flybys during a typical Cassini tour will occur at altitudes of 3000 km or less, and these are considered useful for the radar instrument. Figure 13-9 shows a trace of these flyby ground tracks on a map of Titan. While the Titan flyby geometry must be chosen primarily to produce the desired effect on the Saturn-centered orbit and to allow a free-return to Titan, it is likely that future tour iterations will focus more closely on optimized Titan surface coverage. There will be some trade-off as a Titan flyby is perturbed away from the optimum gravity-assist parameters to a geometry that maximizes Titan flyby science.

Figure 13-9. Titan ground tracks (altitudes below 3000 kilometers).



The baseline tour includes two close flybys of Iapetus and one each of Enceladus and Dione, and a total of 26 other "non-targeted" or accidental encounters within 100,000 km. The velocities of the icy satellite flybys range from about 3 km/sec for the targeted Iapetus flybys to over 20 km/sec for flybys of Mimas and Enceladus near the end of the tour. There are no close flybys of either Hyperion or Phoebe in the baseline tour. The non-targeted flybys sometimes occur only hours or minutes apart on any given orbit, so not all of the flybys will yield useful science.

Cassini Spacecraft System

Saturn Orbiter

The Saturn Orbiter will be the second in the series of Mariner Mark II (MMII) spacecraft, to be used for missions beyond the inner solar system. In keeping with the MMII design philosophy, these will be modular spacecraft that can be inexpensively reconfigured for different missions. Use of proven designs will be maximized; new technologies will be used only where cost effective and only if a viable backup is available. Data interfaces will use standard protocols and standard telemetry formats to minimize the impact of changes. Large margins in mass, power, and performance will be maintained to aid in cost control.

The Cassini spacecraft is illustrated in figure 13-10. At the top of the spacecraft is a 3.67-m diameter high-gain antenna (HGA) with a low-gain antenna (LGA) mounted at the top center. Below the HGA is a 10-bay bus where most of the electronics are stored.

The propulsion module subsystem (PMS) is dominated by two large conispherical bipropellant tanks. These will be of the same design as the tanks to be supplied for the Comet Rendezvous Asteroid Flyby Mission by the Federal Republic of Germany. Not visible in figure 13-10 are the small monopropellant tank which holds hydrazine for the reaction control system, the pressurant tank, and the 1.4-m radio antenna for the Titan probe data relay.

The spacecraft has three deployable appendages: two science booms and one boom for the radioisotope thermoelectric generators (RTGs), which will provide 600 watts of power at the beginning of the Mission. The magnetometer boom canister is located above the RTGs.

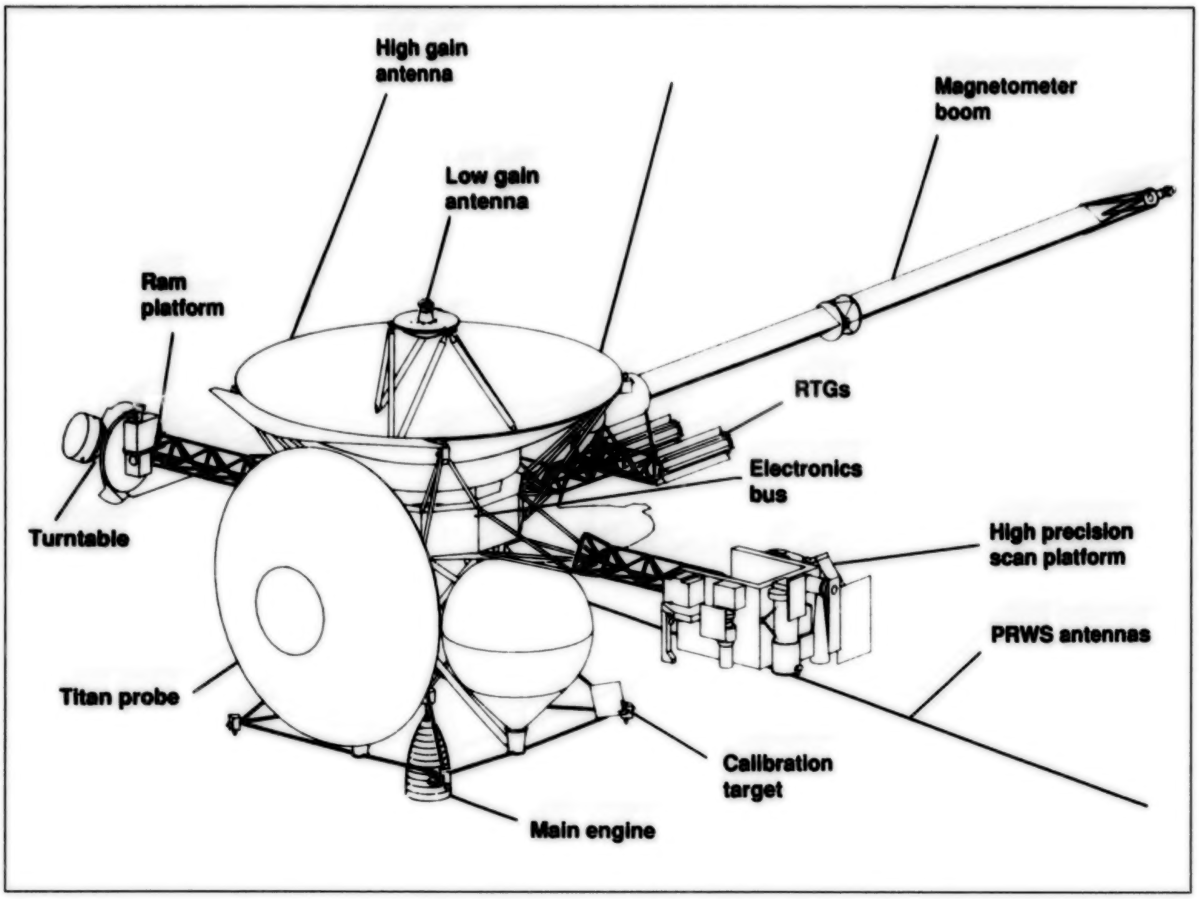

High gain antenna
Magnetometer boom
Low gain antenna
Ram platform
RTGs
Electronics bus
Turntable
High precision scan platform
PRWS antennas
Titan probe
Calibration target
Main engine

Figure 13-10. Cassini spacecraft configuration.

The high precision scan platform (HPSP) will provide two-axis pointing for the remote sensing instruments to an accuracy in inertial space of at least 2.0 milliradians, with pointing knowledge of 1.0 milliradians. The inertial reference unit and the star tracker will also be on the HPSP to maximize HPSP pointing accuracy. The HPSP motion is momentum compensated with reaction wheels mounted on the PMS. The turntable, located on a boom opposite the HPSP, will support the dust and particle analyzing instruments and will rotate continuously at 1 rpm. Also mounted on the turntable boom will be a platform to allow the aeronomy instruments to point in the direction of motion as the Titan flybys skim the upper atmosphere.

Transmission to and reception from Earth will be achieved with the 3.67-m diameter HGA (Voyager design), two LGAs, redundant 10.6-watt solid state amplifiers, Telemetry Modulation Units, transponders, and Command Detector Units. An LGA will be used for both uplink reception and downlink transmission during periods when the HGA cannot point at Earth due to solar constraints and main engine burns, and in the event of an emergency during which HGA pointing is lost. Coverage on the ground will use the NASA Deep Space Network.

The three-axis stabilized spacecraft will be guided by a redundant inertial reference unit known as a Fiber Optics Rotation Sensor, which provides a gyroscopic function with no moving parts and which operates in conjunction with a star tracker and a sun sensor. The attitude control subsystem will control the pointing of the three science platforms and compensate for their motion, and will control the HGA pointing using a redundant reaction wheel assembly. In addition, the subsystem will control both the 400-N main engine gimbals and the 0.2-N reaction thrusters.

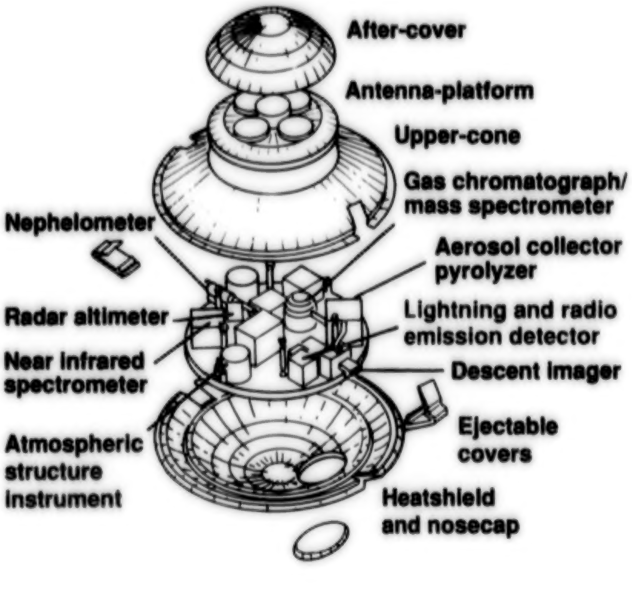
Two digital tape recorders (DTR) with a capacity of 1.8 gigabits each are on the baseline Cassini spacecraft, as are redundant solid-state buffers with a total capacity of 25 megabits. These random-access buffer memories will be used primarily to match data rates between the science instruments and the DTR, as well as between the DTR and the downlink.

Titan Probe

The Probe is currently subject to an industrial phase A study conducted by a European Industrial Consortium. This study will conclude in September 1988. A baseline system design has been developed to demonstrate that all science and mission objectives can be satisfied by a vehicle that will meet the Mariner Mark II interface requirements.

The probe is a blunt-nosed conical body with a half-cone nose angle of 60° and a nose radius of 1250 mm. The aerodynamic decelerator forms the afterpart of the thermal protection system and has an outer diameter of 3.1 meters. The aftercover protects the instruments from radiative heat transfer during descent; the decelerator and the aftercover will be jettisoned when the main parachute deploys. The heat shield is designed to tolerate an entry speed of not more than 7.1 km/sec. Figure 13-11 shows a sketch of the probe; not all details of this figure represent the latest configuration, but the general design concept is similar.

Figure 13-11. Titan probe configuration.


After-cover
Antenna-platform
Upper-cone
Gas chromatograph/ mass spectrometer
Nephelometer
Aerosol collector pyrolyzer
Radar altimeter
Lightning and radio emission detector
Near infrared spectrometer
Descent imager
Ejectable covers
Atmospheric structure instrument
Heatshield and nosecap

Conclusions

The Cassini Mission will address fundamental questions about the formation and evolution of the solar system, pre-biological organic chemistry, plasma physics, atmospheric dynamics, and virtually every other aspect of space science. Together with the Galileo Mission to Jupiter, it represents a systematic and comprehensive approach to the study of the outer planets.

Cassini is now moving toward a launch in 1996. As part of the Mariner Mark II program, Cassini is coupled with the Comet Rendezvous Asteroid Flyby Mission. This allows a tremendous reduction in total costs for these two missions; together they provide complementary solutions to many critical pieces of a very complex puzzle. Cassini is a vital component of an exciting and coherent program of solar system exploration.

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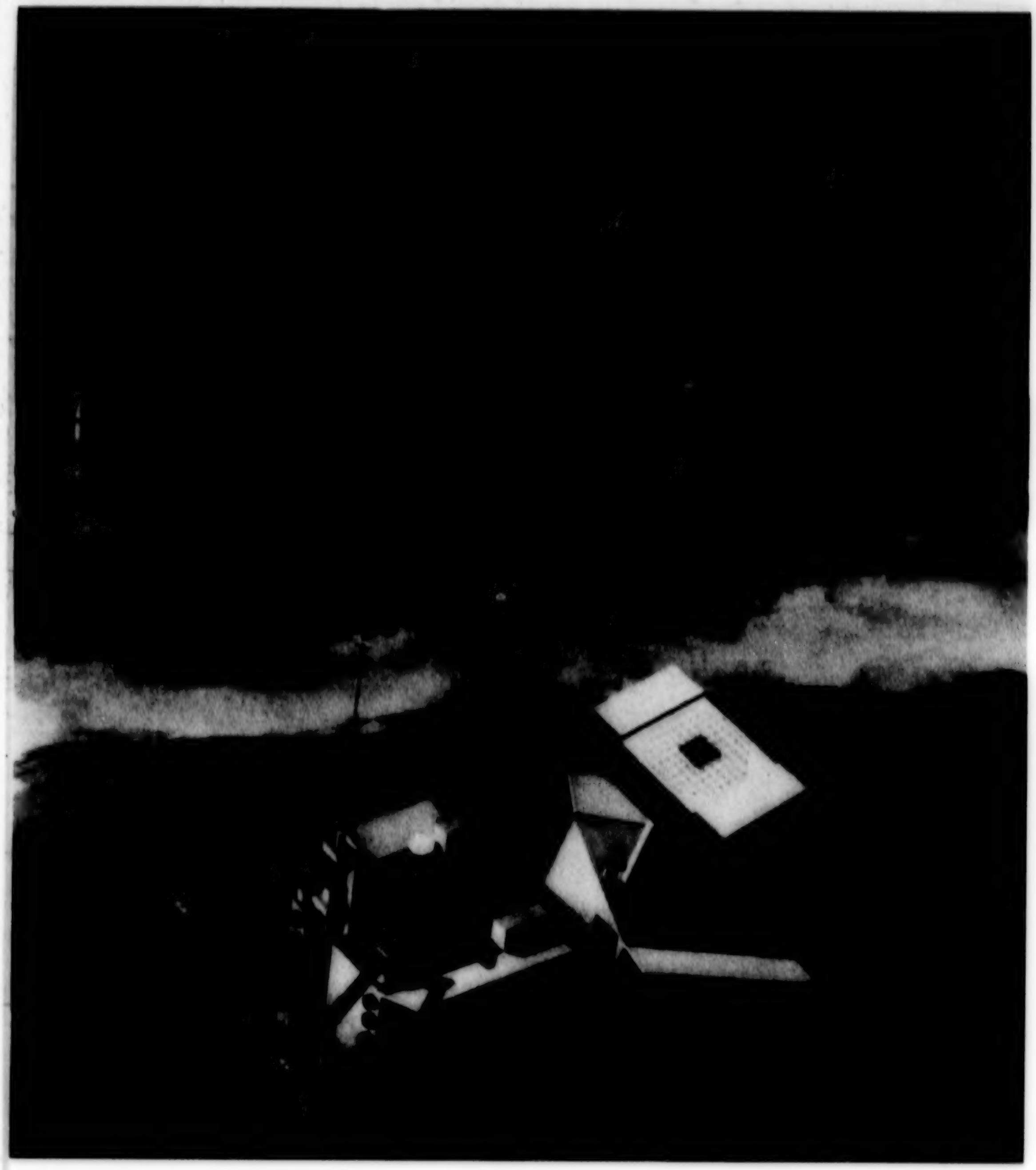

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MRSR: Rationale for a Mars Rover/Sample Return Mission

N93-18560

M. H. Carr

The Solar System Exploration Committee of the NASA Advisory Council has recommended that a Mars Rover/Sample Return mission be launched before the year 2000. The recommendation is consistent with the science objectives as outlined by the National Academy of Sciences committees on Planetary and Lunar Exploration, and Planetary Biology and Chemical Evolution. Interest has also focused on Mars Rover/Sample Return (MRSR) missions, because of their crucial role as precursors for human exploration. As a result of this consensus among the advisory groups, a study of an MRSR mission began early in 1987. The study has the following goals: (1) to assess the technical feasibility of the mission; (2) to converge on two or three options for the general architecture of the mission; (3) to determine what new

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technologies need to be developed in order to implement the mission; (4) to define the different options sufficiently well that preliminary cost estimates can be made; and (5) to better define the science requirements. This chapter briefly describes Mars Rover/Sample Return missions that were examined in the late 1980s. These missions generally include a large (1000 kg) rover and return of over 5 kg of sample. More recently other concepts have been explored that utilize small (10's kg) rovers and return only small (~500 gm) samples. These later studies are not described.

Rationale for Mars

Mars is of special interest among the planets—first, because of what it may reveal about how terrestrial planets with atmospheres formed and evolved, second, because of the possibility that some form of life may have started there, and third, because Mars is certain to be the first planet other than the Earth to be explored by humans. The terrestrial planets are, of course, particularly important for what they might reveal about the Earth. In this respect Mercury, because it is inert and has no atmosphere, is of limited usefulness. Venus is Earth-like in many ways but it is difficult to study because of its thick, opaque atmosphere. Mars, however, is readily accessible to observation. Moreover, it is an active planet that has experienced many of the geological and meteorological processes that have occurred here on Earth. Mars is thus a natural laboratory where processes such as volcanism, crustal deformation, fluvial erosion, and atmospheric circulation can be studied where they have operated under very different conditions from those under which they operate on Earth.

Much of what we know about the formation of the planets is derived from meteorites and from the composition of the Earth and the Moon. A major question is whether the solar system was a relatively violent or a relatively quiet place toward the end of planetary accretion. The "violent" hypothesis suggests that, at that time, there were many Moon-sized objects in eccentric orbits, colliding with one another. In contrast, the "quiet" hypothesis suggests that the inner solar system was occupied by the accreting planets together with small debris in near-concentric orbits. The violent hypothesis is supported by the angular momenta and obliquities of the planets, by the probable origin of the Moon by a large impact, and by the apparent inhomogeneous accretion of the Earth. The quiet scenario is supported by evidence of poor mixing in the early solar system as indicated by radial zonation in the types of asteroids, by variations in the oxygen isotopes between meteorites and the Earth/Moon system, and by differences in noble gas abundances between various objects in the inner solar system.

Samples from Mars will be a major help in resolving this problem. The composition of the upper mantle can be inferred from surface rocks or measured directly in ultramafic inclusions. If Mars' upper mantle has a higher proportion of siderophile elements than is expected if it were in equilibrium with the core, or if the crust is in chemical disequilibrium with the rest of the planet, then inhomogeneous accretion would be indicated and the violent hypothesis would be supported. Conversely, oxygen isotope and noble gas abundance patterns that are unique to Mars would favor the quiet hypothesis.

While we have considerable knowledge concerning the evolution of the surface of Mars, the structure of the interior and its thermal and dynamic state are almost completely unknown. Clarifying these issues will be a major goal of the Rover/Sample Return mission. Of particular interest are the size and composition of the core, when it formed, and how its formation affected the composition of the mantle. On Earth, the relatively high content of siderophile elements in the upper mantle suggests that core formation occurred early, before accretion was complete. Is this also true for Mars? In addition, because Mars lacks plate tectonics, it must dissipate its internal heat in a way different from the Earth, possibly with "hot spot" volcanism playing a more prominent role. The precise mechanism has probably affected the thickness of the lithosphere; the location, nature, and timing of volcanism; and the development of gross, planet-wide anisotropies. Lack of plate tectonics has probably also minimized mixing of mantle and crustal materials so that if a volatile-rich veneer accumulated late, as has been suggested for the Earth, then it is better preserved on Mars. Related key questions are as follows: What is the composition of the crust and how are crustal materials distributed? What is the thickness of the lithosphere, and how has the thickness varied with location and time? What is the current heat flow and are there areal variations? These are all questions that can be addressed by a combination of returned samples, a well-instrumented rover, and long-lived packages of instruments that a rover could deploy.

Features preserved on the Martian surface suggest that the planet has had a long and varied geologic history. The planet has a large-scale asymmetry, having most of the younger surfaces on one hemisphere and most of the older surfaces on the other. Volcanism appears to have occurred throughout the planet's history, possibly up to the present day. Deformation has occurred on both local and regional scales. The surface has also preserved a long record of impacts, despite having been extensively modified by wind, water, and ice. While many of the processes that have shaped the planet are familiar to us on Earth, the results on Mars are spectacularly different. Huge volcanoes have been constructed atop broad regional bulges on the surface. Extensive fault systems suggest a remarkable stability in the state of stress of the lithosphere. Vast canyons have formed. Ice appears to have caused pervasive modification of the surface at high latitudes, and the surface has been subject to episodic floods of enormous magnitude. Why do familiar geologic processes have such large-scale effects on Mars? In what ways do these various processes differ from and in what ways are they similar to geologic processes that occur here on Earth? What has been the nature and sequence of the events that resulted in the configuration of the surface that we see today? Answers to these and similar questions have been, so far, frustrated by lack of compositional, mineralogical, and age information on the surface materials. With appropriate samples in hand we could determine the composition of the crust and the nature of any hemispheric differences. From the planet's seismic activity, the structure of the crust, and the composition of volcanic rocks, we will be able to better understand Martian volcanic processes and how they differ from those on Earth. Seismic and electromagnetic sounding of the surface will also enable us to detect ice and water at depths.

Clarification of the role of water in the evolution of the Martian surface is of special importance. Water plays a crucial role in many geologic processes, including both deep-seated magmatic processes and surficial changes such as weathering, erosion, transportation, and deposition. It is also, of course, essential for any biology. Because most of the water is believed to have outgassed early, the near-surface inventory provides clues about the accretion of the planet, the thermal and fractionation history of the interior, and the formation of the atmosphere. The amount of water also has climatological implications which in turn affect the probability that life started on the planet. There may even be practical implications, for water is a potential source of fuel and sustenance for future piloted missions.

We know that water has extensively modified the surface, both as ice and liquid, but estimates of the amount of water outgassed, and retained close to the surface, range over two orders of magnitude. If the higher figures are correct, Mars may have had warm, wet conditions hospitable to life early in its history. If the lower figures are correct, warm climates could never have prevailed. An MRSR mission should provide a variety of data to resolve this issue. Samples of earlier atmospheres should be preserved in certain rocks, such as impact breccias. Abundant carbonates and other salts near the surface would indicate a large volatile inventory and moist conditions in the past. In addition, water-lain sediments and volcanic products that result from interaction with water and ice should be present if climatic conditions had formerly been warmer and wetter.

The formation, evolution, and dynamics of Mars' atmosphere are closely coupled to the role of water. The formation of the Earth's atmosphere appears to have occurred very early, possibly during global heating as the core formed at the end of accretion. The Mars atmosphere may have formed in a similar way. However, subsequent to the initial outgassing, the planet may have lost significant fractions of its initial volatile inventory as a result of a variety of processes including blow-off by large impacts, thermal escape from the top of the atmosphere, and photochemical escape. In addition, some of the volatiles, particularly CO_2, N_2, and H_2O, became fixed in the ground as carbonates, nitrates, and ices. The climatic history of the planet depends both on the initial inventory of volatiles and on the timing and magnitude of all the subsequent events. Clues concerning these issues will be present in the composition of the atmosphere, in the detailed chemistry of unweathered, igneous rocks, in the products of weathering at different times in the planet's history, and in the variety of past depositional environments. Data in all these areas should be acquired by an MRSR mission.

Mars, of course, is of special interest because of the possibility that Martian life may have once existed. The probability that living forms may still be present on Mars is very small. The reasons for doubting that life is currently present have been previously summarized. They are (1) the absence in the surface materials of the reduced carbon and nitrogen compounds that living forms are expected to produce; (2) the lack of liquid water, which is required for almost all biochemical reactions; (3) the low temperatures; (4) the high surface flux of ultraviolet (UV) radiation which tends to break down complex organic molecules; (5) the absence of gas disequilibria in the atmosphere; and (6) the adequate chemical explanations for the three Viking biology experiments. It has, nevertheless, been argued that the negative results at the Viking landing sites do not necessarily rule out survival of life forms in local, more hospitable environments elsewhere on the planet. However, warm, wet locations protected from the UV radiation, and with energy sources that could be utilized by living forms, must be very rare, if they exist at all. Moreover, because of the general hostility of the planet to life and the lack of a global, interconnected biota, colonization of local niches is unlikely. For all these reasons, the chances for finding present-day life on Mars are considered so small that it has not been included as one of the primary objectives of the MRSR mission being studied for the late 1990s. More definitive experiments to re-examine the possibility of present-day life on Mars could, however, be an objective of follow-on MRSR missions.

While the chances of extant living forms existing on Mars are extremely small, the probability of life developing early in the planet's history is quite different. Fluvial erosion of ancient surfaces and the presence of bedded sediments of possible lacustrine origin show that water was at least intermittently present on the Martian surface at times in the past. The most spectacular fluvial features on the planet are those that formed as a result of catastrophic floods. Possible causes are volcanic melting of ground ice or release of groundwater under high pressure. Spectacular as these features are, they are thought to have little implication for past climates. Indeed, they could have formed under the present very cold conditions which are hostile to life. The much smaller branching valley networks are, however, suggestive of former, warmer climatic conditions. They appear to have formed by the slow erosion of running water, much like terrestrial river valleys, and not by sudden catastrophic release. For such slow erosion to occur, temperatures at the surface probably had to be close to or higher than freezing; otherwise the streams would freeze. The valley networks are almost all on very old surfaces. This suggests that warm conditions prevailed mostly early in the planet's history.

Greenhouse calculations indicate that atmospheric pressure had to be at least 1 atmosphere to raise the mean surface temperatures to 273 K. Both warmer climatic conditions and a thicker atmosphere have, therefore, been suggested for early Mars, that is, for the period during, and shortly after, the early heavy bombardment by meteorites. During this early period the planet's interior was also warmer and the planet could have had a magnetic field, which would have protected the surface from the adverse effects of solar wind and solar flares, as in the case of the Earth. Indeed, conditions on Mars and Earth may have been similar prior to 3.5 billion years ago when life on Earth originated. Because of these considerations, the search for prior life is one of the major scientific goals of the MRSR mission.

A final reason for interest in Mars is that it is a likely site for future piloted missions. The MRSR mission is expected to be launched around the end of the century. At that time piloted Mars missions may be only a decade or two away and plans for such a mission could be well advanced. Before humans go to Mars we will need to know considerably more about its environment so that instruments, life support systems, return vehicles, and so forth can be relied upon to function at Mars for long periods of time. The omnipresent dust is of special concern since it might be carried into the spacecraft to contaminate the life-support systems and it might be ingested by the astronauts. In addition, various technologies necessary for piloted missions need to be tested at Mars before risking their use on human missions. Additional objectives of the MRSR mission will be, therefore, to gather environmental information and to test key technologies necessary for the eventual human exploration of Mars.

Rationale for Samples

Experience gained with the Moon emphasizes the enormous power of returned samples when placed in the context of other global data. Sample data are the basis for almost all of our current ideas on how the Moon formed, what fractionation events subsequently occurred, how the highlands and maria differ in origin, and when all the various events that led to the present configuration took place. The degree to which this understanding could have been achieved solely by remote analysis is a matter of conjecture, but there is ample reason to be skeptical. Having lunar samples in hand allowed the complete analytical and intellectual capacity of the scientific community to focus on the problem of the Moon's evolution. Instead of having a small pre-determined set of analytical techniques applied to samples, the approach could be both all-encompassing and flexible, with the analytical emphasis shifting as the meaning of each set of results became better appreciated. There is no reason to believe that these enormous advantages should be any less for Mars. Indeed, the apparently more complex geology on Mars and the biologic implications enhance the importance of the kind of comprehensive examination that a returned sample allows.

One of the main advantages of returned samples is that they permit the use of analytical techniques that are too complicated to be performed remotely. Among the most important are absolute dating methods, such as Rb-Sr, K- Ar, Sm-Nd, and U-Th-Pb. These fix the whole chronology of the planet's evolution and, in addition, provide essential information on fractionation mechanisms, on the formation of the crust, and on the extent to which surface materials are recycled. Knowledge of trace elements to the parts per million to parts per billion level is necessary for determining the bulk composition of the planet and for understanding a wide variety of processes of chemical differentiation. Examples of the value of trace elements are the use of rare earth elements in elucidating early events on the Moon and the use of iridium in understanding events at the Cretaceous-Tertiary boundary on Earth. Such determinations are difficult and strain the capabilities of laboratories on Earth. The only practical way to do them for Mars is on returned samples. Petrographic examination, while routine on Earth and commonly conducted in conjunction with other techniques such as microprobe analysis, is also very difficult to perform remotely because of the difficulties of sample preparation and the large amount of human judgment needed in the analysis. However, these techniques are necessary for identifying rock types, interpreting their origin, and recognizing secondary alteration. Similarly, returned samples are essential for the location and determination of complex organic compounds and for microscopic studies needed in the search for evidence of former life.

Apart from the difficulties of remote analysis, there are methodological advantages to having samples returned to Earth for analysis. Perhaps the most serious disadvantage to *in situ* analysis is its inflexibility. Planetary missions have long lead times so that the set of analytical techniques to be used must be chosen at least 10 years ahead of analysis. Once the choice is made it cannot be changed as our perception of the planet changes or when the first analytical results are acquired. As with Viking, it may take 20 years to respond to a new set of results in a program based solely on remote observations. In contrast, with samples in hand, different techniques can be applied as suggested by each set of new analytical results and new techniques can be developed if needed. Returned samples can also be preserved to be repeatedly re-examined as our ideas about the planet's evolution mature or as new techniques are developed. Finally, should we ever send humans to Mars, samples will be available to test for any potentially harmful effects that the reactive Martian materials may have on humans.

Rationale for a Rover

A Rover would have three main functions: (1) to support the acquisition of samples, (2) to perform *in situ* analyses, and (3) to explore. Support for sampling is needed to ensure that the samples returned to Earth are representative of the planet's variety. To accomplish this the Rover must be able to move; to have the appropriate analytical capabilities to recognize differences in the accessible materials; and to have the appropriate tools for sampling, packaging, and delivering to the vehicle samples that are to return to Earth. The Viking results emphasize the importance of mobility. Both landers set down on fine-grained debris. While there were rocks within reach, all the rock-like chips analyzed were highly weathered and were almost certainly clods of the local weathered materials rather than unweathered rocks. The debris is probably highly weathered and not comminuted material, with abundant fragments of unaltered rock as on the Moon. The material is thus of limited usefulness in understanding the evolution of the planet. Moreover, spectral characteristics of the rocks within the Viking fields of view suggest that no more than two types are present. Thus, a sample-gathering vehicle at the two Viking landing sites would have to move to get any rock samples, and to move beyond the field of view to get a variegated set.

Selection of appropriate landing sites will be crucial for the success of the MRSR mission. To address a wide range of science questions we need a variety of samples such as volcanic rocks of old, intermediate, and young ages, ancient impact breccias, water-lain sediments, polar-layered deposits, soils, wind-blown dust, and ices. Landing sites with more than two major rock units exposed will be difficult to find. Moreover, the choice of site will inevitably be a compromise between the desire for safe landing and the desire to go to scientifically variegated locations to address a variety of science objectives. The mobility of a Rover both increases the chances of getting a variegated sample set and alleviates the conflict between the safety and science desires. With mobility, the landings could all be in safe, relatively uniform terrains, while the sampling is mostly done in nearby, variegated, more scientifically interesting locations.

The Rover must have capabilities in addition to mobility to support sample acquisition. It must, for example, be able to recognize differences in the materials available for sampling to guard against return of numerous samples of the same materials. This will require some remote sensing ability to indicate where different types of samples may be available and some sample analysis capability to identify more precisely the differences between samples. Such analytical capability should be directed at both the high-temperature major rock components and the low-temperature, more volatile rich components, that are of climatological and biological interest. Instruments necessary to do a first-order characterization of the area to be sampled should also be available so that sampling strategy can be guided by knowledge of the local variety in the available materials. Finally, the Rover must have instruments appropriate for documenting the location and geologic context of the sample site so that the subsequent analytical results can be appropriately interpreted. How far the Rover has to roam and how capable a set of instruments is needed to support sampling remain to be determined.

Biology places particularly stringent needs on site selection and sample acquisition. First we wish to sample at locations where former biologic activity is most likely. Prime candidates are ancient lake deposits since lakes would have provided the conditions necessary both to support living organisms and to preserve evidence of their presence. In many areas of the ancient highlands, valley networks converge on local basins. Lakes are likely to have existed in these areas. Unfortunately, these low-lying areas are also preferred sites for accumulation of volcanic deposits and identification of areas where sediments are accessible and not covered with volcanics will not be easy. If such areas are found and landed on then one is confronted with the task of selecting the most promising sample. We are likely to be able to return to Earth only 5-10 kilograms so every sample is precious. We are unlikely to be so lucky as to find direct visual evidence for life, such as stromatolite colonies. If life ever started on Mars the most likely means of detecting it will be through chemical evidence. But organics tend to be destroyed at the surface so we need to acquire samples that have spent much of Mars' history below the surface. Sampling debris from a relatively young impact crater in the lake sediments is one way of achieving this. In addition, we want to be able to do some preliminary testing for organics and possibly isotopic fractionation so that the samples that we bring back to Earth are the ones most likely to preserve biologic evidence.

In situ analyses are important in their own right, independent of sample acquisition. The COMPLEX report on exploration of the inner planets is very careful to emphasize the importance of both returned samples and *in situ* analyses. Certain characteristics of a planet are not susceptible to study by means of samples. The most obvious are the structure and dynamics of the planet's interior and the structure and general circulation of the atmosphere. In addition, some properties of the surface materials must be measured *in situ* because they change in response to local seasonal and diurnal changes in the environment. The extent to which such measurements will be part of an MRSR will depend on the capabilities of the various vehicles involved, and that remains to be determined. Most previous studies have recommended inclusion of various geophysical and meteorological objectives in the goals of an MRSR mission. Experiments such as active and passive seismometry, heat flow determination, and monitoring the atmosphere are, therefore, candidates for inclusion in the mission. The geophysical and meteorological instruments could be at a fixed location or deployed at various places by the Rover.

The third function of the Rover is to explore. At present we have detailed knowledge of the surface of Mars only at the two Viking landing sites, so having a Rover on the surface will greatly extend our knowledge of Mars at ground level. This capability is important for understanding the processes that control the small-scale morphology of the surface and for understanding the environment in which human explorers will ultimately have to work. This attribute of the mission is also the one most readily appreciated by non-scientists. The

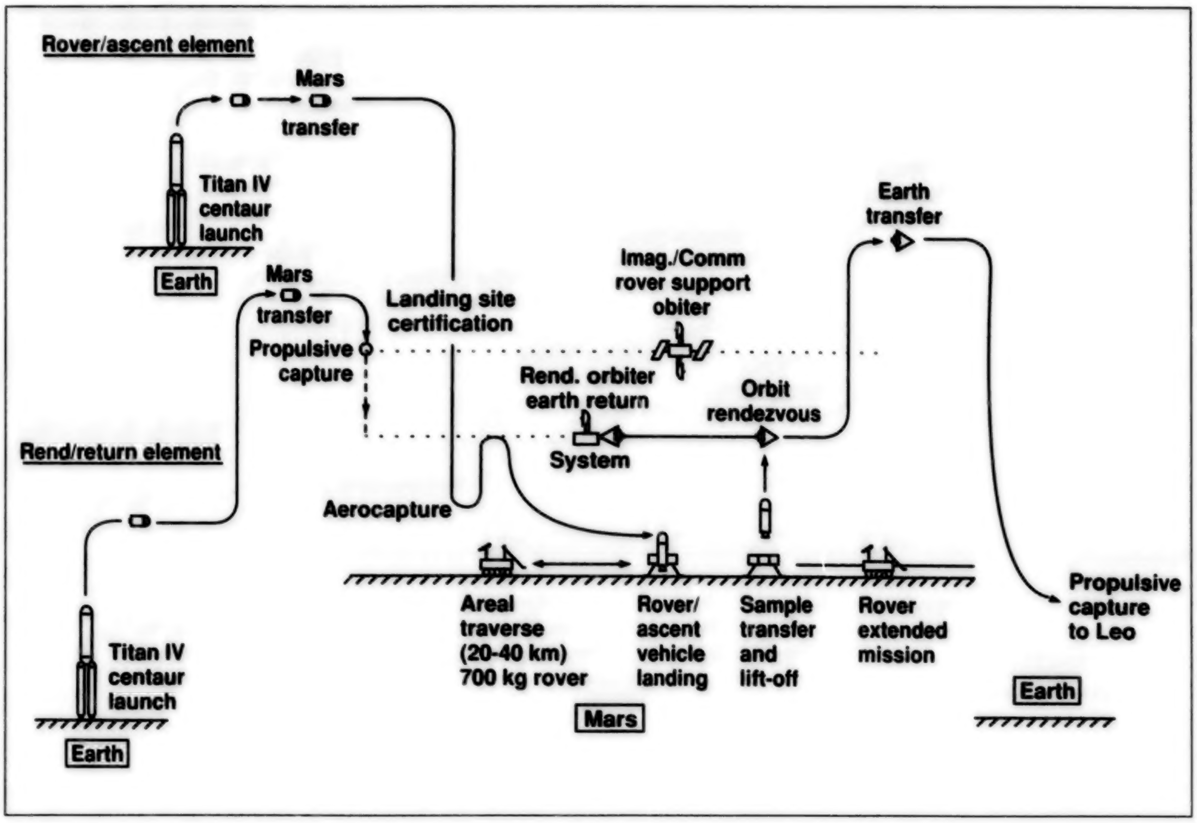
Rover/ascent element
Mars
transfer
Titan IV centaur launch
Earth
Mars
transfer
Landing site certification
Imag./Comm rover support obiter
Earth transfer
Propulsive capture
Rend. orbiter earth return
Orbit rendezvous
System
Rend/return element
Aerocapture
Areal traverse (20-40 km) 700 kg rover
Rover/ ascent vehicle landing
Sample transfer and lift-off
Rover extended mission
Propulsive capture to Leo
Titan IV centaur launch
Earth
Mars
Earth

Figure 14-1. Possible scenario for a Mars Rover-Sample Return mission.

main exploration requirements are high mobility and imaging. The three functions—sample acquisition, *in situ* analysis, and exploration—place different demands on the Rover. What emphasis should be placed on the different functions is being debated.

Possible Mission Scenarios

An MRSR mission would involve several vehicles, which can be packaged for launch to Mars in different ways. Figure 14-1 shows one possible mission scenario. Most of the scenarios considered in the late 1980s involved two separate launches for each mission. However, if the nation developed a high lift vehicle then the mission could be accomplished with a single launch. In the configuration shown in figure 14-1, the Earth-Return Vehicle for the samples and the Orbiter that is used to support ground operations at Mars are combined in the first launch; the Rover and Mars Ascent Vehicle are combined in the second launch. This combination has the advantage that the Orbiter can be sent to Mars well ahead of the Rover to certify and characterize the landing site. In other possible combinations, the Rover and Ascent Vehicle are launched separately and land separately at Mars. While this has advantages in terms of launch mass margin, it has the disadvantage that the Rover and Ascent Vehicle must rendezvous on Mars. In the scenario in figure 14-1 the Rover and Ascent Vehicle land together, so rendezvous is not a problem.

After a trip to Mars of approximately 1 year, the Earth Return Vehicle and the Orbiter will be injected into Mars orbit. The Orbiter will then proceed to examine the preselected landing sites to make sure they are safe for landing and to provide a detailed map of the landing area for Rover operations. Meanwhile the Rover/Ascent-Vehicle combination will have been placed into Mars orbit using aerocapture to conserve fuel. Once the landing site has been judged safe, the Rover and Ascent Vehicle will descend to the surface probably using a combination of aeromaneuvering, deceleration by parachute, and finally, terminal propulsion. After landing, the Rover will be free to separate from the Ascent Vehicle and perform its mission.

How mobile the Rover will be, what analytical capability it will have, and how sophisticated its sampling tools will be are currently all undefined. Missions of different complexities are being studied from those in which the Rover is tethered to the Ascent Vehicle and can rove only 100 meters from it, to missions in which the Rover can traverse distances of several tens of kilometers. In all the scenarios the Rover makes looped traverses, returning repeatedly to the Ascent Vehicle to deliver samples. The Rover is expected to have a remote sensing capability to navigate and to recognize compositional variations in the surrounding scene. It will also have tools necessary to collect loose rocks, to sample bedrock and large boulders, and to extract samples from 1-2 meters below the surface. The Rover should also have the capability of analyzing major elements and volatile compounds in samples and nearby materials and to deploy geophysical/meteorology stations. After several months on the surface the samples that have been collected will be launched from the Martian surface into Mars orbit. Once the samples have been launched, the Rover will continue exploring the surface.

After launch, the Ascent Vehicle will rendezvous with the Earth-Return Vehicle, in orbit around Mars, and transfer to it the sample capsule. The Earth-Return Vehicle will then be injected into the Earth transfer orbit for the trip back to Earth which will take approximately 1 year. At Earth approach various options are possible. A sample capsule could be released for direct entry into the Earth's atmosphere or the sample capsule could be placed into Earth orbit either propulsively or by aerocapture. What procedure is followed will depend strongly on how much emphasis is placed on back contamination issues.

More recently, ways of implementing sample returns have been examined that are very different from that just described. Miniaturization of spacecraft components, rovers and analytical instruments, together with advances in analytical techniques that permit comprehensive analysis of very small samples, has stimulated the design of much smaller, and substantially lower cost sample return missions. It may be feasible, for example, to launch a sample return mission with a miniature rover on a Delta launch vehicle. This being so, a different exploration strategy might be followed in which sample return missions of limited capability are sent to many sites rather than emplacing a substantial capability at few sites, as described above.

Conclusions

In 1992 the United States will place in orbit around Mars a vehicle designed to systematically map chemical, mineralogical, and topographic variations of the surface and to monitor atmospheric activity for a full Martian year. This will be followed in 1994 and 1996 by a Soviet mission consisting of rovers and balloons. Both U.S. and Soviet advisory groups agree that the next step should be a Rover/Sample Return mission to the planet. Such a mission could be launched in the late 1990s. It would address a broad range of planetological and biological issues and is expected to have an impact on Mars science comparable to the impact that Apollo samples had on lunar science. Major objectives of the mission are to determine whether life ever started on Mars and what conditions were like on Mars very early in its history when life began on Earth. The mission is also expected to pave the way for human exploration by acquiring key environmental data about Mars and testing technologies necessary to place humans on the planet and return them safely to Earth.

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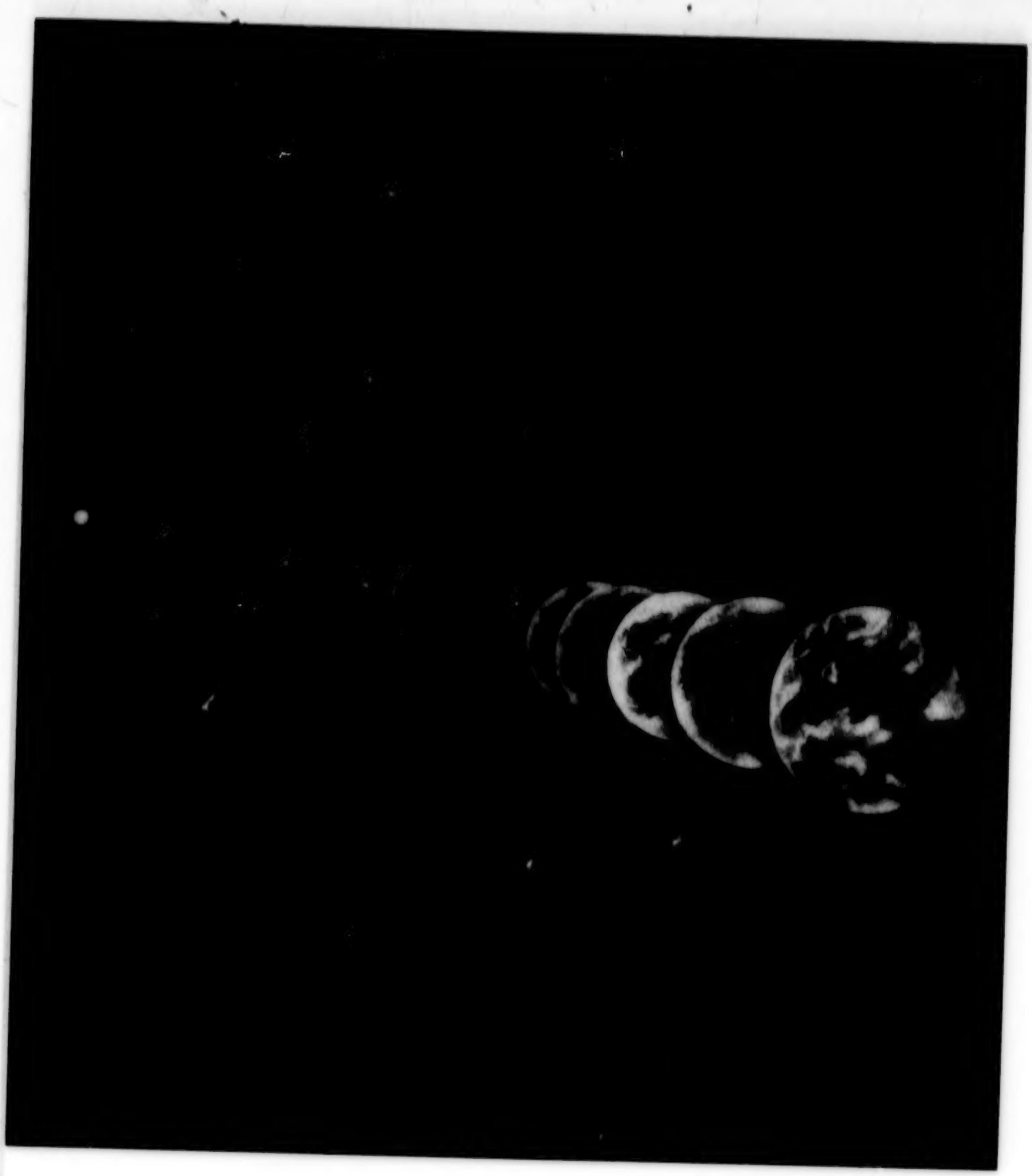

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Exobiology: The NASA Program

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Exobiology is an interdisciplinary program of scientific research supported by the Life Sciences Division of NASA. As its goal, the program seeks to increase knowledge of the origin, evolution, and distribution of life in the universe: answers are sought as to how the development of the solar system led to a habitable planet, how life originated on Earth, what factors influenced the course of biological evolution, and where else life may be found in the universe. These questions have broad scientific and cultural significance by addressing the possibility that life is unique to Earth while investigating prospects for its existence and detection elsewhere. A general theory for the natural origin and evolution of living systems within the context of the origin of the universe may eventually arise from knowledge gained in the systematic search for answers to these age-old questions.

NASA is chartered to promote "the expansion of human knowledge of phenomena in ... space" (National Aeronautics and Space Act of 1958). The rationale for conducting the Exobiology Program has been, and continues to be, firmly rooted in the agency's charter. Our understanding of biology and the natural history of life on Earth strongly indicates that biological evolution is subject to the vicissitudes of planetary and solar system evolution and that life could well arise and evolve on suitable planets. For these reasons, unparalleled opportunities to contribute to the state of knowledge of the biological sciences are embodied within the missions and projects associated with NASA's exploration of space.

At the scientific core of the Exobiology Program lie issues concerning the natural history of life in the universe, all of which can be addressed most fruitfully in the context of evolution. In the galactic, cosmic, solar system, planetary, or environmental sense, "evolution" refers to the course of change over time as a consequence of the changing thermodynamic state of the universe. It is within this milieu that life arose and evolved in concert with the physiochemical universe. From a biological point of view, evolutionary change in living systems is wrought by physical, environmental, and biological events: mutation, reproduction, and natural selection. Life, then, may be viewed as a unique product of countless changes in the form of primordial matter—ultimately changes wrought by the processes of astrophysical, cosmochemical, geological, and biological evolution, which are all aspects of the evolution of the universe. Therefore, the historical relationship between the two courses of evolution and the history of processes involved in the dynamics of that relationship constitute the unifying conceptual framework of the Exobiology Program.

Recognition of the concept of evolutionary interplay has led to a focused program of scientific research on how the natural history of that interplay took place. The Solar System Exploration Committee (SSEC) of the NASA Advisory Council has articulated several goals as NASA's mission capabilities expand and diversify. Three of the committee's stated goals are

1. To determine the origin, evolution, and present state of the solar system.

2. To understand Earth through comparative planetology studies.

3. To understand the relationship between the chemical and physical evolution of the solar system and the appearance of life.

The thinking of the SSEC and other advisory committees, along with the present state of space life sciences and NASA's schedule of missions and projects for the upcoming years, have been factored into the formulation of goals and objectives for the Exobiology Program. As stated before, the overarching goal of the program is to understand the origin, evolution, and distribution of life in the universe. To do this, the Exobiology Program seeks to provide a critical framework and some key research to allow NASA to bear the combined talents and capabilities of the agency and the scientific community, and the unique opportunities afforded by space exploration. To provide structure and direction to the quest for answers, the Exobiology Program has instituted a comprehensive research program divided into four program elements which are being implemented at several of NASA's research centers and in the university community. These program elements correspond to the four major epochs in the evolution of living systems:

1. Cosmic evolution of the biogenic compounds;

2. Prebiotic evolution;

3. Origin and early evolution of life; and

4. Evolution of advanced life.

The overall research program is designed to trace the pathways leading from the origin of the universe through the major epochs in the story of life.

Cosmic Evolution of the Biogenic Compounds

The first epoch, the cosmic evolution of the biogenic compounds, encompasses galactic time and distance scales; it begins with the nucleosynthesis of the biogenic elements, which make up the bulk of all living materials, and ends with the transformation of their compounds into planetesimals, the building blocks of planets. Within these bounds are included relevant phenomena in stars, interstellar clouds, presolar nebulae, and the primitive objects in the solar system. These phenomena precede the first stage of planetary evolution: this epoch of cosmic evolution sets the stage for the formation and evolution of the solar system and its planetary bodies.

The goal of this program element is to determine the history of the biogenic elements (C, H, N, O, P, and S) from their birth in stars to their incorporation into planetary bodies.

NASA's development of astronomical telescopes for use in space, its program for the study of pristine objects like comets and asteroids (and materials derived from them) using ground-based laboratories and spacecraft, all provide a framework of opportunities to enhance our knowledge of the events in this epoch. In the late 1990s, research facilities on Space Station Freedom will allow exobiology experiments to be conducted under microgravity conditions. Of particular interest to exobiologists is the Cosmic Dust Collection Facility, planned as an attached payload, and the Gas-Grain Simulation Facility, planned for use within the pressurized volume of Space Station Freedom.

Cosmic Evolution of the Biogenic Elements

Supporting NASA Missions/Projects

- Past and Present
    - Kuiper Airborne Observatory
    - International Halley Watch
    - ER-2 IDP Collection Flights
    - Ground-Based Radio Astronomy
    - Meteorite Collection Programs
    - Infrared Telescope Facility
    - Ames Vertical Gas Gun
- Future
    - Comet Rendezvous Asteroid Flyby
    - SOFIA
    - SIRTF
    - SSF Cosmic Dust Collector
    - SSF Gas-Grain Simulation Facility
    - Comet Nucleus Sample Return

Figure 15-1.

Investigations and activities conducted within this program element continue to increase our understanding of this epoch (fig. 15-1). All of the biogenic elements have now been detected in space. Laboratory data compared with astronomical observations of interstellar clouds have revealed the presence of molecular ices in the protostellar environment and these ices have been found to contain such biologically significant materials as water, ammonia, carbon monoxide, alcohols, nitrates, aldehydes, and other organic molecules. The implication is that the chemical constituents needed for the development of life are available throughout the universe.

Organics have also been detected in comets such as Halley. Comets are objects believed to have been preserved intact, below their surfaces, since the first epochs of solar system formation. Analysis of the Murchison meteorite revealed the presence of amino acids with a deuterium/hydrogen ratio similar to that in interstellar space, providing a link in the incorporation of interstellar materials into our solar system. Similarly, diamond residue from the Allende CV3 meteorite revealed information on temperatures and pressures encountered in the presolar nebula and affecting the formation of the biogenic elements. Thus, it is becoming apparent that chemistry in space may be linked directly to the chemistry we see on Earth.

Prebiotic Evolution

The second epoch, prebiotic evolution, begins with the formation of planets and ends with the emergence of living organisms from the prebiotic milieu. The research in this area includes studies of two kinds of processes: planetary and molecular. Planetary processes occur on global or local scales as a consequence of the development of planets and are responsible for setting the physical and chemical conditions of the planets' various environments wherein living systems may arise. Molecular processes may also occur on macro- or micro-environmental scales, but they comprise the chemical and physical mechanisms by which the structures and functions that are attributable to primitive life forms develop and assemble in response to energy sources. This differentiation has been made for purposes of the programmatic organization; research on the existence of a continuum between the two—and their essential interdependence—is acknowledged.

The goal of this program element is to understand the pathways and processes leading from the origin of the planet to the origin of life. This involves the investigation of the planetary and molecular processes that set the physical and chemical conditions within which chemical evolution occurred and living systems arose.

From the standpoint of the origin of life, prebiotic evolution is the most important epoch and the one about which the least is known, primarily because of a lack of an extant fossil geological record on Earth and because of the immaturity of theories dealing with origins of both planets and biological systems. This area continues to be targeted for major emphasis (fig. 15-2). NASA's capabilities are uniquely suited to enhancing present knowledge of prebiotic evolution, insofar as the agency's program of planetary exploration will allow important gaps in the geological record to be filled by studies of other planets and small bodies in the solar system.

Prebiotic Evolution

Supporting Missions/Projects

- Past and Present
  - Apollo
  - Viking
  - Pioneer
  - Mariner
  - Meteorite Collection Programs (USARP)
  - Deep-Sea Vent Explorations (NOAA)
- Future
  - Mars Observer
  - Titan-Cassini Mission
  - Mars Rover/Sample Return
  - Hubble Space Telescope
  - Astrometric Telescope/CIT

Figure 15-2.

The studies conducted should explore a continuum of planetary possibilities, including bodies devoid of organic chemicals, those conceivably undergoing (or having undergone) organic chemical evolution, and those possibly harboring life, extant or extinct. Carefully selected sites for exploration may reveal the remnants of chemical precursors to living systems and allow examination of cases where planetary evolution may have diverged from the path that led to the origin of life on Earth. Lifeless planets, meanwhile, provide examples of environments where prebiotic chemical evolution or subsequent biological evolution never occurred or ended. The search for other planetary systems is an integral element of these studies and extends the search for life and potentially habitable environments beyond the solar system.

Evolution of Early Life

The third and fourth epochs encompass biological evolution, beginning with the emergence of living systems from the prebiotic planetary environment and carrying through to the appearance of intelligence in advanced organisms. The goal of research into the early evolution of life is to determine the nature of the most primitive organisms, the environment in which they evolved, and the way in which they influenced that environment. Research in the early evolution of life focuses on deciphering the records of biological evolution preserved in rocks and in contemporary organisms and ecosystems, with a view toward understanding how the evolution of early microbial life was influenced by changes in surface environmental conditions on early Earth. A further useful division of this research is made into studies of the geological record and the biological record, insofar as different strategies, methods, and materials are employed in each. Figure 15-3 illustrates former and planned future missions supporting this area of research.

Evolution of Early Life

Supporting
Missions/Projects

- Former and Current
    - Viking Missions
    - Deep-Sea Vent Explorations (NOAA)
- Future
    - Mars Rover/Sample Return
    - EOS (Baseline Studies)
    - VLBI Studies of Extrasolar System Planets

Figure 15-3.

A major question for exobiologists in the 1990s and beyond is that of the planets and life. Fossils 3.5 billion years old have been located in the Warawoona Group in Western Australia. These are the earliest records of life found thus far. Their discovery has pushed back temporal estimates for the origin of life millions of years, to a time when the environments of early Earth and early Mars may have been much more similar than they are now.

Indications are that both planets may have been suitable for development of life during this remote epoch. Since we know that life evolved on our own planet, Earth, the question arises, Did life evolve on the planet Mars during this more clement period, and if not, why not?

Current exobiological research using the lakes in the dry valleys of Antarctica as analogs of possible Martian paleolakes is providing insight into ways to conduct the future search for remnants of life on Mars. Additionally, studies of endolithic organisms found in these dry valley regions may offer clues for future searches of bio-markers of past life on the arid red planet.

Evolution of Advanced Life

The research associated with the study of the evolution of advanced life seeks to determine the extrinsic factors influencing the development of advanced life and its potential distribution. This research includes an evaluation of the effect of extraterrestrial influences on the appearance and evolution of multicellular life, conducted by (1) tracing the effects of major changes in the Earth's environment on the evolution of complex life, especially during mass extinction events; (2) determining the effects of global events and events originating in space on the production of environmental changes that affected the evolution of advanced life; and (3) by searching for evidence of advanced life elsewhere in the galaxy. Planned future missions, as well as previous missions, supporting the study of evolution of advanced life are listed in figure 15-4.

Evolution of
Advanced Life

Supporting
Missions/Projects

- Former and Current
  - Apollo Missions
  - Viking Missions
  - Pioneer Missions
  - Mariner Missions
- Future
  - SETI Microwave Observing Project
  - The Great Observatories

Figure 15-4.

Preparation for Future Missions

It is hoped that during the next 10 years the Exobiology Program will fly experiments on each of the following missions described, as the opportunities arise.

(1) Mars Observer. A mission that will determine the global elemental and mineralogical character of the surface of the planet Mars.

(2) The Search for Extraterrestrial Intelligence (SETI). A multichannel receiver and spectrum analyzer have been developed and tested. The next step is a systematic search using existing ground-based radio telescopes.

(3) Earth Orbiting Telescopes. An active effort is ongoing in the astrophysical community to develop and fly Earth-orbiting platforms capable of observations in wavelength intervals and at resolutions unattainable from the surface of Earth. Especially noteworthy opportunities include the Hubble Space Telescope and the future Space Infrared Telescope Facility (SIRTF). Astronomical observations of key interest to exobiology include: the detection of interstellar organic molecules, the spectroscopic characterization of the carbonaceous components of interstellar and circumstellar dust, and the detection of protoplanetary nebula and planetary systems.

(4) Comet Rendezvous-Asteroid Flyby (CRAF). This mission has been recently canceled. If resurrected, it will make a close rendezvous with a comet nucleus. It will make measurements to determine the nature and characteristics of a comet nucleus, thereby providing exobiologists with much needed information of chemical evolution in the outer solar system.

(5) Cassini. Its objective is to characterize the composition and structure of Titan's atmosphere and of the Saturnian system. There is a considerable amount of methane on Titan, and Voyager detected at least nine organic compounds more complex than methane. The atmospheric haze observed is believed to be composed of solid organic material. It seems certain that Titan will reveal much about the organic chemistry that can occur in an abiological setting. Current plans for exobiology involvement on a Titan probe center on analyses of the organic compounds in the gas phase and in the solid aerosol material.

(6) Cosmic Dust Collection Facility. Designed for Space Station Freedom, this device will enable the nondestructive capture of high-speed extraterrestrial particles in the micrometer size range. As particles are captured, a complete description of their state of motion will be obtained. This will allow for determination of their point of origin and hence their astrogeophysical source. Capture methods, such as the use of underdense collection media, also offer near-term capabilities that can be exploited on spacecraft before Space Station Freedom is operational. Grains that originate from known comets and grains that are interstellar in origin would be of particular importance.

(7) Space Station Freedom Gas-Grain Simulation Facility. Space Station Freedom will provide unique opportunities to study physical and chemical processes in microgravity. Simulation studies designed to investigate the circumstellar, interstellar and protosolar nebular environments and processes can provide new information on topics such as grain condensation, comet mantle formation, grain-grain and gas-grain interactions involving the biogenic elements and compounds.

(8) Mars Environmental Survey Network Mission (MESUR). This mission set will deploy a series of penetrators and/or small stations to several sites on the surface of Mars. Capabilities envisioned include descent imaging, seismometry, and geochemical analysis of surface and near-surface materials.

(9) Mars Rover/Sample Return. This mission will involve detailed study of specific sites on Mars. (Sites will probably be selected on the basis of data returned by the Mars Observer and Network Missions.) Anticipated here are

opportunities to study the history of the biogenic elements, principally water, on Mars and to conduct a more intensive search for traces of past organic activity and remnants of past life.

Mission opportunities of relevance to exobiology will continue to arise in the early 21st century as the space program matures and its capabilities expand. These include: Comet Sample Return, Asteroid Sample Return, Large Deployable Reflector (an Earth-orbiting telescope), Pluto flyby, Neptune orbiter, Lunar Base, and Mars Base.

Emphasis will most likely be placed on the sample return missions to Mars and to a comet, the two augmented missions recommended by the SSEC. These targets are accessible with technology available in the near future; collecting devices and associated hardware will be able to survive and function on the surfaces of the bodies and previous missions will have provided enough information about the targets to allow detailed planning. Both missions are of prime importance to the Exobiology Program.

Summary

The goal of the Exobiology Program is to understand the origin, evolution, and distribution of life throughout the universe. Because of recent advances in diverse contributing fields of science, the Exobiology Program is now ready to advance the theory that life is a natural consequence of the physical and chemical processes that comprise the evolution of the cosmos, and to embark upon a scientifically rigorous course of action to test this theory. It is felt that by synthesizing the results from the studies described, it should be possible to generate a set of rules that describes the path of life from nucleosynthesis of biogenic elements in stars through evolution of planetary systems to the origin of life and evolution of intelligent life.

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Future NASA Solar System Exploration Activities: A Framework for International Cooperation

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The Solar System Exploration Program today is in a critical transition between a glorious past and an uncertain, but potentially exciting, future. Planetary exploration has moved from the breathtaking pace of the 1960s and 1970s, when successful, groundbreaking missions were launched one after the other, and it is entering an era of fewer, but far more sophisticated, missions. The late 1980s and early 1990s are crucial times for solar system exploration, and the years between now and the end of the 20th century are filled with both challenges and opportunities.

Both the Challenger accident on January 28, 1986, and the subsequent cancellation of the Shuttle-Centaur upper stage drastically affected the planetary program. All the missions under development—Magellan, Galileo, Ulysses, and Mars Observer—have been delayed. These delays have forced the program to divert money and resources originally intended to support research and new program development into the continued planning and engineering of grounded missions. The unexpected absence of an energetic Shuttle-Centaur makes it necessary to carry out complex gravity-assist trajectories—which, in turn, produce much longer trip times—to reach the outer planets.

At the same time, other nations have begun major efforts in planetary exploration. The increasing presence of non-U.S. spacecraft beyond Earth orbit was especially noticeable in 1986, when a 5-spacecraft "Halley Armada" was launched to study the famous comet. The Armada consisted of the European Space Agency's highly successful Giotto Spacecraft, Japan's Planet-A and Sakigake, and the U.S.S.R.'s two VEGA Spacecraft. Although NASA provided critical ground support to the Armada with its Deep Space Network, no U.S. spacecraft was sent to join it. In addition, the Commonwealth of Independent States (C.I.S.), formerly the U.S.S.R., is now actively planning a decade of ambitious efforts to explore Mars. This effort has already begun with the launch of the Phobos Mission in July 1988, to study the moon of Mars of the same name, and it may lead to an unmanned sample return around the year 2000.

Clearly, things are different from the "Golden Age of Planetary Exploration" that took place during the last 25 years. For several reasons, future solar system exploration requires a new approach. First, we must recover from all the changes, good and bad, that have occurred since 1980. We must build our greatly increased understanding of the solar system into a series of more capable and more sophisticated exploration missions. We must develop the new potential for international cooperation that has become available. And we must begin to view planetary missions as something new—not just a way of obtaining valuable exploration for the sake of science, but also as the means to obtain essential data that we need to pursue a new national goal—expanding human presence beyond Earth orbit and into the solar system itself.

Plans and Outcomes: The Solar System Exploration Committee and What Happened

In 1980, upon recognizing that planetary exploration had changed radically, NASA formed the Solar System Exploration Committee (SSEC) of the NASA Advisory Council to begin a fundamental review of the entire planetary program. The charter of the SSEC was to formulate a long-term program for planetary exploration, within a constrained budget, that would ensure a vital, exciting, and scientifically valuable effort through the turn of the century. The most urgent priority was to provide stability to the program while ensuring its continued progress. The SSEC, therefore, first formulated a Core Program of planetary exploration that was stable, produced exciting and important scientific results, made the maximum use of existing facilities and technologies, and could be carried out within a reasonable level of continued funding.

The Core Program was developed around a number of elements. An initial sequence of Core Program Missions included the Venus Radar Mapper (now renamed Magellan), a Mars Geoscience/ Climatology Orbiter (now Mars Observer), a Comet Rendezvous/Asteroid Flyby, and a Titan Probe/Radar Mapper. Within the Core Program, a Planetary Observer series was to be established, consisting of low-cost, modestly scaled missions within the inner solar system. For missions beyond the inner solar system, the Mariner Mark II series was to be developed--a straightforward, modular spacecraft that could easily be adapted to a variety of missions to the outer solar system and to primitive bodies—comets and asteroids. A common mission operations and information system was to be developed for all missions after Magellan. Finally, the Planetary Research and Analysis program was to be significantly strengthened, to expand our ability to analyze currently available data, to perform necessary ground-based research, and to develop instrumentation for the missions of the Core Program (table 16-1).

Initial progress toward establishing the Core Program was encouraging. Congressional approval of the Magellan Mission as a "new start" in 1983 was followed closely by approval for the Mars Observer (the first Planetary Observer) in 1984, but subsequent progress has been slow. Congress has endorsed the Planetary Observer concept, but the Mariner Mark II series still awaits adoption. The Comet Rendezvous/Asteroid Flyby (CRAF) Mission, the first of the Mariner Mark II missions, was submitted for, but did not receive, new start approval in fiscal years 1988 and 1989. However, the inclusion of this mission—together with the proposed Cassini Mission to Saturn—in the FY 1992 budget is a hopeful and encouraging sign for the future. However, to ensure that a robust Solar System Exploration Program still exists at the turn of the century, these missions must be approved soon. Finally, NASA has implemented the multi-mission operations concept, but the research and analysis activities still remain funded at an undesirably small fraction of the total program budget.

Since the SSEC made its recommendations, other events and new realities have drastically altered the environment in which solar system exploration must make its plans. Some of the events have produced major changes, both positive and negative, in our technical ability to explore the planets during the next few years. Other, less tangible, events have changed the political and social climate in which the national space program, and its planetary exploration component, will operate in the future.

In 1984, President Reagan endorsed a national commitment to human presence in the solar system through the initiation of the Space Station Program. Since then, Space Station Freedom has become an important factor in planning for future planetary exploration, and one that was not anticipated by the SSEC. Despite the fact that the Space Station will be physically located in Earth orbit, it can provide important capabilities for several important and exciting planetary exploration projects, two of which are being studied. The Cosmic Dust Collection Facility would trap particles of cosmic dust, measure their velocities, and determine their trajectories; the particles would then be returned to Earth for detailed laboratory analysis. The Astrometric Telescope Facility would begin a decades-long program of precise measurements of the motions of nearby stars to detect the possible presence of large planets around them.

Table 16-1: Planetary Missions Through Year 2000

Approved Missions

Missions	Target(s)	Launch	Mission Highlights
Magellan	Venus	April 1989	Radar mapping from near-polar, highly elliptical orbit for one Venusian year. Ninety percent of the surface mapped.
Voyager 2	Neptune	August 1989	Voyager 2 encountered Neptune and is now off to study the interstellar medium.
Galileo	Jupiter	October 1989	Asteroid flyby en route; using a corkscrew trajectory that will use gravity assists from Venus and Earth, Galileo will reach Jupiter in late 1995. An instrumented probe will enter the Jovian atmosphere; the orbiter will operate for 20 months, successively approaching most of Jupiter's large satellites.
Ulysses	Sun	October 1990	Ulysses will study the Sun out of the ecliptic plane.
Mars Observer	Mars	September 1992	Mapping of the Martian surface for one Martian year. Data collected on climatology, surface composition, topography, gravity field, and magnetic field.

Proposed Missions

Missions	Target(s)	Launch	Mission Highlights
Lunar Observer	Moon	mid-1990s	Extended orbital study of the lunar surface and gravitational and magnetic fields.
Comet Rendezvous Asteroid Flyby (CRAF)	Asteroid Hamburga Comet Kopff	August 1995	Asteroid flyby followed by comet rendezvous in 2000. Spacecraft will fly in formation with Kopff for 3 years as comet approaches and passes through perihelion. CRAF will deliver a penetrator to study the interior of a comet's nucleus.
Cassini	Saturn	April 1996	Fly past the asteroid Maja in 1997, gain gravity-assist from Jupiter, and arrive at Saturn in 2002 for 4 years of study of the Saturnian System. Probe will be ejected into Titan's atmosphere.
Mars Rover/ Sample Return	Mars	late 1990s	Advanced mission to land on Mars, traverse its surface, and return 5 kilograms of samples for study on Earth.

Future modifications of the Space Station might support more ambitious planetary exploration projects. Preliminary studies are being made to define Space Station-attached orbital telescopes for full-time planetary observations in both visible and infrared wavelengths. More complex Space Station developments could include the assembly, checkout, and launch of large planetary missions or the preliminary quarantine and analysis of returned extraterrestrial samples in a special laboratory module. These ideas will be developed in detail as plans for large planetary missions, and for the Space Station, evolve together in the future.

The loss of Challenger in 1986 had far-reaching effects on the Solar System Exploration Program. The restoration of the Space Shuttle to flight status was critical for near-term planetary exploration, and the successful launch of Discovery on September 29, 1988, was an important achievement for planetary exploration as it was for all of NASA.

By heroic efforts, the four missions under development at the time of the Challenger accident—Galileo, Ulysses, Magellan, and Mars Observer—have been preserved, but all of them have been delayed, and the increased costs and necessary mission redesigns have severely strained the planetary program. The Space Shuttle launches of Magellan in April 1989, and of Galileo in October 1989, were critical milestones along the path to a viable and stable planetary program for the future.

Launch plans for later missions are more flexible as a result of NASA's decision to proceed with a "mixed-fleet" policy, in which space missions would be launched either by the Space Shuttle or by expendable launch vehicles. When this plan is in effect, it will provide a substantially more robust launch capability for planetary missions.

The Solar System Exploration Program is following the establishment of this policy with extreme interest. As an initial step, the Mars Observer was designed to be compatible with either the Space Shuttle or an existing Titan-3 expendable; in late 1988, NASA decided to launch Mars Observer in 1992 on the Titan-3, rather than on the Shuttle, as was originally planned. A more powerful launch vehicle, the Titan-4, is expected to be available possibly for CRAF or other planetary missions in the 1990s.

Another factor in planning the future of solar system exploration is that attention is increasingly turning toward two relatively unexplored areas: the outer solar system and the primitive bodies. A huge amount of new information about the outer planets and comets was obtained from the Voyager 2 encounter with Uranus in January 1986, and from the International Halley Armada's flybys of Halley's comet in March of that year. In August 1989, Voyager 2 encountered Neptune and now continues on to study interstellar space. Also in 1989, Galileo was launched and is en route to study Jupiter and its moons. The next proposed major mission, CRAF, will visit an asteroid and a comet, and CRAF will be followed closely by the Cassini Mission to Saturn and its moon Titan. These less-explored areas of our solar system present intriguing possibilities for understanding the origin of the solar system, the formation of planets, and the origin of life.

Outside the Solar System Exploration Division, even outside of NASA, events have been occurring that will strongly influence the future of planetary exploration. These events have produced significant changes in the emphasis of our nation's space program, and these changes must be accommodated in the Solar System Exploration Program as well.

In 1986, the President's National Commission on Space released its report, "Pioneering the Space Frontier." This report planned for a very ambitious program for the future, emphasizing lunar bases and human exploration of Mars as long-term national goals. A year later, an internal NASA report to the Administrator by scientist and former astronaut Sally Ride, titled "Leadership and America's Future in Space," was released. This report recommended four major new initiatives for the long-term future of NASA: global Earth studies (or the "Mission to Planet Earth"), the Solar System Exploration Initiative (which includes expanded planetary exploration and a series of Mars Sample Return missions), a human outpost on the Moon, and the human exploration of Mars. At the same time, NASA established a separate office, the Office of Exploration (OEXP), to study human missions to the Moon and Mars and to determine the requirements for precursor missions to establish the data base needed for these explorations.

A new National Space Policy was signed in 1988, endorsing the maintenance of national leadership in space and explicitly including a new goal to "expand human presence and activity beyond Earth orbit into the solar system."

These events have combined to place the planetary exploration program in a far different situation from that envisioned by the SSEC 5 years ago. Major, and generally successful, efforts are being made to recover from the loss of the Space Shuttle and the Centaur stage and to launch our current missions. The return of the Space Shuttle to flight status was a heartening development, since both Magellan and Galileo used the Shuttle for their 1989 launches. The use of expendable launch vehicles for future planetary missions is being actively considered, and a decision has already been made to launch Mars Observer on a Titan-3 in 1992. Plans are also being made to use the unanticipated capabilities of Space Station Freedom (not considered by the SSEC) for new kinds of planetary exploration from Earth orbit. Especially important is the new interest in long-term human exploration of the Moon and Mars. This development has given a new and dual significance to certain planetary missions, not only as a means of scientific exploration, but also as precursors for future human landings.

Goals and Approaches for the Future*

The fundamental goals and approaches for planetary exploration as stated and embraced by the SSEC have not been changed by later events. These goals and approaches are the outcome of long-standing and continuing interactions between NASA and its advisory committees, and they are still accepted by NASA and the Solar System Exploration Division as the basis for the U.S. planetary program. Slightly revised from previous descriptions, these goals are

1. Origin and Evolution–To determine the present nature of the solar system, its planets, and its primitive bodies, to understand how the solar system and its objects formed, evolved, and (in at least one case) produced environments that could sustain life.

2. Comparative Planetology–To better understand the planet Earth by determining the general processes that govern all planetary development and by understanding why the "terrestrial" planets of the solar system are so different from each other.

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*At the time this paper was first drafted, August 1988, the information contained herein was current. As four years have passed since that time, many changes to NASA's Solar System Exploration Program have taken and will continue to take place.

3. Pathfinders to Space–
To establish the scientific and technical data base required for undertaking major human endeavors in space, by carrying out essential precursor activities which include the survey of near-Earth resources, the characterization of planetary surfaces, and the search for life on other planets.

The manner in which programs and missions are carried out to reach these goals will also vary little from the guidelines established in the past. In particular, the planetary program of the 1990s will stress the five traditional themes described below.

1. Program Balance–
Planetary exploration will continue the balance of activity that has yielded such exciting results in the past. Two kinds of balance are involved: balanced activities between different planetary science disciplines, and balanced exploration of the three fundamental classes of solar system objects: terrestrial planets, primitive bodies (comets and asteroids), and outer planets, or gas giants.

2. Progressive Series of Missions–
The exploration of other worlds via spacecraft will continue the rational sequence of increasingly sophisticated and scientifically detailed missions that has been carried out in the past. This sequence progresses from space reconnaissance, which is accomplished by fast flybys, through exploration, which is achieved by orbiters and atmospheric probes, and into intensive study, which consists of landers, rovers, and sample returns.

3. Strong Ground-Based Support–
The mission activities, although more visible, depend on an essential foundation of ground-based research, mission operations, instrument development, and advanced studies. This component of the planetary program must be maintained, and in some cases, augmented.

4. International Cooperation–
Cooperating with other nations and groups of nations in space exploration has always been a part of NASA's activities. But such cooperation is especially important for the future, because the capabilities and plans of other nations have increased. A special emphasis on U.S.-U.S.S.R. activities was established by the agreements on space cooperation between the two countries. Cooperative studies and other activities, especially involving the future exploration of Mars, continue with the C.I.S.

5. Interdisciplinary Cooperation–
Because we have learned so much about the universe in the past 25 years, it is now possible to address larger and more specific scientific questions, many of which extend across the narrow boundaries of the scientific disciplines established in the early days of the space program. Such currently exciting problems as the relations between the Sun and interplanetary plasmas, the combined formation of stars and solar systems, the search for gravity waves, and the origin of life now involve more than one of the Divisions in NASA's Office of Space Science and Applications. As we enter a new period of space exploration, in which missions will be both less numerous and more sophisticated, the Solar System Exploration Division will work closely with other disciplines to ensure that new missions address this wide array of questions as completely as possible.

As we prepare to meet our long-standing goals in both traditional and new ways, we must assess the current status of planetary exploration. Before the launch of Magellan and Galileo in 1989, more than ten years had passed since the last U.S. planetary mission, Pioneer Venus, was launched in 1978. But even with this long hiatus, the achievements of the last 25 years constitute an awesome accomplishment and provide a solid base for new and exciting exploration.

By the end of the 1980s, all the planets except Pluto had been studied by reconnaissance phase (flyby) missions which began with Mariner 2 in 1962, and ended with the Voyager 2 flyby of Neptune in August 1989 (fig. 16-1). More than two dozen planets and moons, some only dots of light even in the best telescopes, have been transformed into a fantastic variety of worlds, some familiar, some bizarre, and all surrounded by exciting problems and new scientific questions. The exploration phase of orbiters and probes has been established for all the inner planets, with the notable exception of Mercury, and the stage is now set to expand the intensive studies phase of the Moon and Mars that was briefly initiated by the Apollo and Viking programs.

Finally, new developments in technology and the establishment of Space Station Freedom (fig. 16-2) will make it possible to attack important planetary problems from Earth orbit—the nature and composition of cosmic dust, the detection of other planetary systems, and the long-term observation of other planets.

One of the most important long-term consequences of the Challenger loss has been the focusing of NASA on the issues of risk, reliability, and what to do to prevent failure. These heightened concerns were most visible during the long process of restoring the Shuttle to flight status, but they have major implications for how planetary exploration will be conducted in the future.

Planetary exploration has always been risky. In the early days, when missions were simple and inexpensive, the problem of risk was addressed in a straightforward way: by constructing and launching two duplicate spacecraft, each with the same payload. If one spacecraft or launch vehicle failed, its twin could carry out the planned mission. This method of risk reduction also provided an important bonus: if the first spacecraft worked perfectly, then the second could be used for a complementary mission.

Figure 16-1. False-color image of Neptune. Red areas are semitransparent haze covering planet.

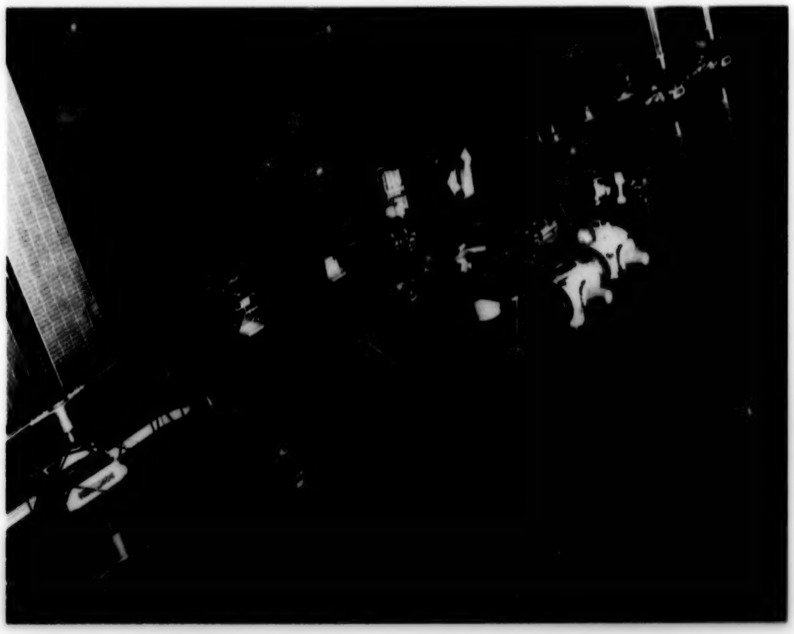


Figure 16-2. Space Station Freedom Concept.

This duplication, despite its obvious advantages, became increasingly difficult to implement as missions grew more complex and expensive, and Voyager was the last mission to which this method could be applied. With Galileo, risk reduction had to be built into the system—redundant systems, complementary instruments, contingency plans for operations, and a high degree of confidence in the launch systems. These precautions, although necessary for a single mission, further increased the cost. In addition, this protection is not complete; because only one mission is involved, a single failure in the spacecraft, the launch vehicle, or even in one or more payload instruments can result in a major loss of data or even the loss of the mission itself.

The planetary program is presently responding to two factors: the heightened NASA-wide concern about risk, and the absence of any series of common missions to provide the originally planned backup. The program has, therefore, developed a risk-reduction approach that is based, not on fully developed backup missions, but on the purchase of essential spare components for both the spacecraft and its payload for any given mission.

These spares will reduce risk in two ways. First, they will be available to support the schedule for assembly and testing of the original mission. If one component has problems during this period, it can be quickly replaced by the spare so that the schedule and the launch opportunity can be met. Second, in the event of a major failure in the original mission, the available spares will provide an essential nucleus for rebuilding the spacecraft and its payload for the next launch opportunity.

This policy is already being put into effect. In FY 1988, Congress appropriated funds to purchase spares of certain critical spacecraft and instrument components for the Mars Observer Mission. These spares will provide increased assurance that the current schedule for Mars Observer, based on a 1992 launch, can be met. In the event that the Mars Observer has a major failure, the spares will be used as the nucleus for reconstructing another mission to meet the next launch opportunity to Mars.

If the Mars Observer Mission is successful, the spares can then be used as the basis for the next Planetary Observer mission, the Lunar Observer, which is under study for a new start in the early 1990s. Once this mission is approved and under way, it will obtain its own set of spares, which can then be used in the next Planetary Observer mission, and so forth. Similarly, the planned development of the CRAF Mission will include similar essential spares, which can then be used in the next Mariner Mark II mission, the Cassini Mission to Saturn and Titan.

No amount of spares will ever substitute for the main goal: making the greatest possible effort to ensure that each mission is fully successful in and of itself. The new spares will support that goal by reducing the risk that minor problems during mission development could produce major delays. And the "rolling spares" philosophy, in which the spares for each mission provide the nucleus for the next one in the series, will provide both support and continuity for planetary exploration in the 1990s.

Origin and Evolution: The Outer Solar System

Clues to the origin and evolution of the planets can be found throughout the solar system. But the most fundamental of these clues, and the best-preserved information, can be found in the bodies that lie in the chill darkness beyond the orbit of Mars.

During most of the 4.6 billion years since its origin, the solar system has been recovering from the violent, dynamic processes that created it. On Earth and the other terrestrial planets virtually all evidence of the earliest epochs has been erased by continuing internal processes—volcanism, mountain-building, and the development of oceans, atmospheres, and life. By contrast, the comets, the asteroids, and the bodies of the outer solar system have largely escaped these processes, and these objects have thus retained some record of early planetary formation.

Because they are small, comets and asteroids have never been able to generate the internal heat that powers large planets and produced wholesale alterations of the materials that formed them.

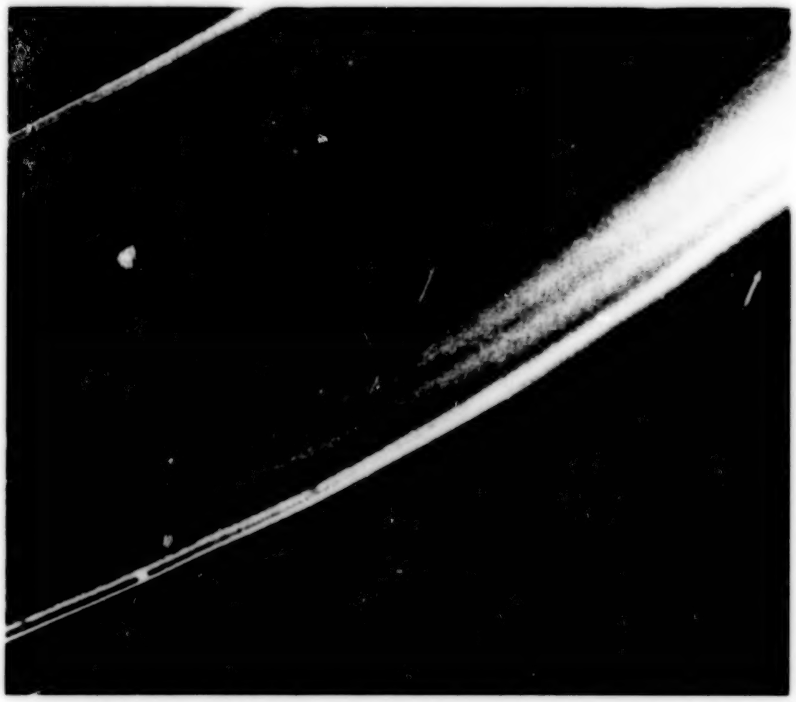
In currently held theories for planetary formation, both the terrestrial planets and the cores of the outer planets grew through the accretion of countless small solid bodies called planetesimals, which were similar to present-day asteroids and comets. One group of these original small bodies has been preserved as asteroids in the asteroid belt, a relatively stable region between the orbits of Mars and Jupiter. Disturbances by Jupiter's gravity probably prevented the asteroids from accreting in a single planet-sized body, and thus they retain many of their primitive chemical and physical characteristics.

Just as asteroids (and the meteorites derived from them) may be samples of the inner parts of the original solar system, so comets may be icy remnants from the accretion of the more distant outer planets. According to this view, the newly formed outer planets then gravitationally scattered the comets to the most remote part of the solar system—the distant Oort cloud, halfway to the nearest star. There the comets remain, in the deep freeze of space, until a passing star, a giant molecular cloud, or the tidal forces of the galaxy perturb them into new orbits that bring them back into the planetary region. An alternative theory is that comets are icy planetesimals that formed in the solar nebula beyond the orbit of Neptune. In either case, comets have most likely been preserved in a condition approaching their original state.

The giant planets of the outer solar system, Jupiter and Saturn in particular, are so massive, and so cold, that they have retained essentially all the material from which they were originally made. Thus, these planets are expected to contain a representative sample of the original solar nebula material, possibly altered to new chemical forms, but retaining the elemental and isotopic signatures of the primordial solar nebula.

The ring systems around the outer planets permit us to study the dynamic interactions of small co-orbiting bodies, which are similar to the interactions among planetesimals that led to the accretion of planets (fig. 16-3). The moons of the outer planets, stunningly revealed by the Voyager pictures, are mostly a variety of cold, frozen worlds that nevertheless record histories that in some ways parallel the evolution of the terrestrial planets. Ranging from highly modified to the nearly pristine worlds, these satellites, like the comets, may contain records of the early solar system, frozen and preserved for our inspection.

Figure 16-3. Backlit view shows continuous distribution of fine particles throughout ring system of Uranus.



The exploration of the outer solar system is an area where American leadership remains unchallenged. Only U.S. spacecraft have ever passed the asteroid belt. Pioneers 10 and 11 and Voyager 1 made historic encounters with Jupiter and Saturn. Voyager 2, after passing Jupiter and Saturn, made the first encounter with Uranus in January 1986, and then met with distant Neptune in August 1989.

Despite these achievements, we have obtained only quick glimpses; we have visited, but we have not yet begun to explore. And although we learned much about comets through data gathered by the Halley flybys, what we learned is only an intriguing glimpse into the mysteries of comets. We have yet to encounter an asteroid, we have yet to rendezvous with a comet and study it for any

Figure 16-4. The Galileo spacecraft, with its large radio antenna furled like an umbrella, is shown in this artist's design. The long boom extending to the right contains sensitive instruments for magnetic measurements.

length of time, and we have yet to place an orbiter around any of the outer planets. To understand these worlds, and to learn what they can tell us about the solar system, we need to explore them at length, and to learn much more about them: what they are like, how they formed, how they change with time, and why they are so different.

We are now ready to begin this exploration. We can now build on what we have learned to design new missions that provide the longer observation times and the more sophisticated studies that we need. We can respond to the challenges of the outer solar system, the excitement of new scientific explorations, and the opportunities to gather new knowledge about the solar system and about ourselves.

The exploration phase for the outer solar system began with the October 1989 launch of the Galileo Mission to Jupiter and its moons (fig. 16-4). When it arrives at Jupiter in 1995, Galileo will begin our first long-term examination of this huge world, its strange moons, and its giant magnetic field.

Galileo is actually two spacecraft which will be launched as a single unit and will then separate shortly before arrival at the planet (fig. 16-5). One spacecraft, a probe, will plunge into Jupiter's atmosphere and will descend on a parachute until it is crushed by the increasing pressure. During its brief lifetime, it

Figure 16-5. In this artist's concept, the Galileo probe plunges into the thick, swirling atmosphere of Jupiter and releases its reddened heat shield to begin its measurements. The Orbiter (upper left) will serve as a data relay for the short-lived probe.

will make the first direct analyses of the upper layers of Jupiter's atmosphere. The second spacecraft is a long-lived orbiter that, for nearly 2 years, will map the behavior of Jupiter's atmosphere, record its magnetic field and radiation belts, and make orbital excursions for closer looks and more detailed studies of the four large moons revealed to the Voyager spacecraft in 1979.

Galileo will also explore on its way to Jupiter. As it passes through the asteroid belt in the early 1990s, it will be directed to pass close to two asteroids, Gaspra and Ida, and it will give us our first close-up view of these tiny bodies. After Galileo, all missions to the outer solar system will make similar flybys of other asteroids.

After the loss of Challenger and the cancellation of the Centaur upper stage, the path to Jupiter required major revisions. Because of the change in launch opportunity and the lower available launch energies, the trip time to Jupiter is now longer than originally planned, and a complicated trajectory, involving multiple gravity-assist flybys at Venus and Earth, is required. Nevertheless, Galileo will reach Jupiter and its moons in 1995. Galileo involved substantial cooperation with Germany in the design of the retro-propulsion module, the rocket motor that will place the spacecraft in orbit around Jupiter.

After Galileo, all missions to the outer solar system will use the Mariner Mark II series of spacecraft envisioned by the SSEC. The Mariner Mark II is a modular spacecraft design, with 3-axis stabilization like Voyager, that can be easily modified to accommodate a variety of missions to the outer solar system, including the primitive bodies. The spacecraft is specially designed to meet the wide range of demands imposed by the nature of these missions to the outer solar system: generating power at great distances from the Sun; transmitting large amounts of data to Earth over vast distances; and providing highly accurate pointing for the precise aiming of penetrators, probes, and remote sensing instruments, as well as for the proper orientation of antennas. Of equal significance is the need for a high degree of autonomous operation and reliability.

The Mariner Mark II missions will focus on exploring the origin and evolution of the solar system by studying the primordial material and the early organic chemistry to be found in the outer planets and small bodies. The Mariner Mark II program begins with a dual initiative: the Comet Rendezvous/Asteroid Flyby (CRAF) Mission and the Cassini Mission to Saturn.

The CRAF Mission is a critical element of the Solar System Exploration Program for two reasons: it initiates the Mariner Mark II program, and it begins the major search for our origins in some of the most primitive solar system objects, the comets and asteroids (fig. 16-6). The molecular and mineralogical compositions of cometary material will yield information about the physical and chemical conditions that existed in the early solar nebula. Studies of the dark material on comets and on carbonaceous asteroids will tell us what these organic compounds are, how they formed, and what role they might have played in the origin of life. Studies of the interactions of the solar wind with the gas and dust emitted from an active comet will reveal the chemical and physical processes that controlled the origin and evolution of dust, gas, and

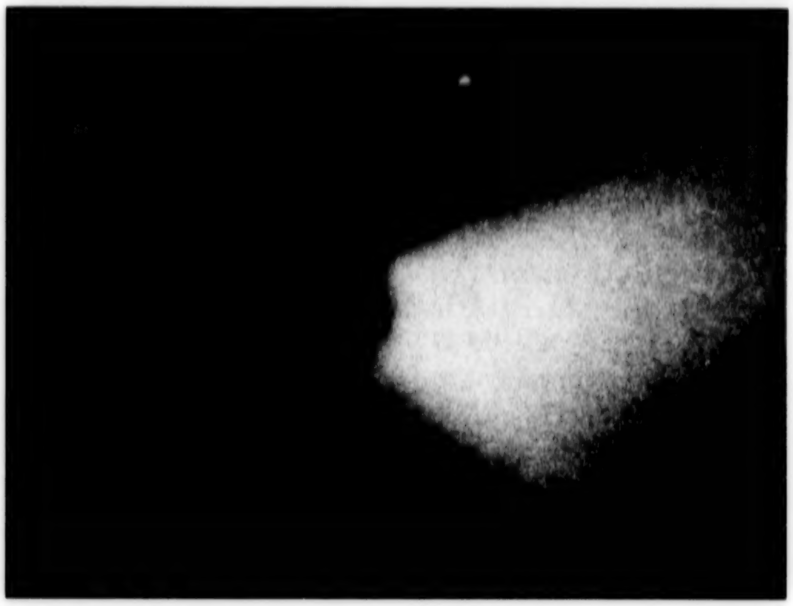


Figure 16-6. Halleys comet, 1986.

small solid planetesimals in the solar nebula.

During its 10-year journey, CRAF will fly past at least one asteroid and will explore a comet in detail. To discover how a comet was formed, and to see how it evaporates under the influence of the Sun's warmth and light, CRAF will rendezvous with a typical short-period comet and will fly in formation with it as it moves around the Sun from its most distant to its closest point of approach. During this multi-year observation, instruments on CRAF will determine the elemental, isotopic, molecular, and mineralogical composition of cometary material and will also investigate the comet's microscopic and large-scale physical structure. As the comet becomes more active near the Sun, CRAF will study the physics and chemistry of the comet's atmosphere and its interaction with the sunlight and the solar wind.

During the asteroid flybys before the comet rendezvous, CRAF will determine the physical properties of one or two asteroids, measuring such unknown qualities as their sizes, shapes, bulk densities, and rotation periods. Instruments trained on the asteroids as the spacecraft flies by them will photograph their surface features and will collect data about their mineral and chemical composition.

Once through the asteroid belt, CRAF will fly on to rendezvous with a comet and to begin long-term observations that will produce a major increase in our knowledge of comets. The Halley Armada flybys of 1986 provided brief *in situ* measurements of cometary gas and dust, and gave us our first glimpse of the dark nucleus of a comet. In many respects, however, the Halley flybys raised more questions than they answered, and the long rendezvous mission designed for CRAF represents a substantial advance.

In a rendezvous mission, a spacecraft follows an orbit around the Sun that precisely matches the comet's orbit. When the spacecraft catches up with the comet, it makes a small rendezvous maneuver, and then the spacecraft and the comet travel together indefinitely. Among the many significant advantages of a rendezvous mission over a flyby are its ability to produce longer and more detailed mapping, the ability to adapt to changing conditions as the mission progresses, and—especially important—the ability to observe the comet's entire lifestyle, from cold inertness in the outer solar system to violent activity near the Sun.

While the comet is quiescent, CRAF will use a wide variety of remote sensing instruments to study it from close range. About a year after CRAF arrives at the comet, the spacecraft will fire a large, needle-like penetrator directly into the surface of the comet's solid nucleus (fig. 16-7). Instruments in the penetrator will measure the chemical composition of the nucleus and will determine the temperature profile within the nucleus, its surface strength, and the exact nature and composition of the mixture of ices, dust, and organic materials that are thought to make up the nucleus. The penetrator, powered by batteries, will transmit its data back to the CRAF spacecraft for about 1 week.

When the comet becomes active near the Sun, the spacecraft will move in and out through its atmosphere, collecting dust for on-board analysis. Several plasma detectors on the spacecraft will study in detail the interactions between the comet's gas and dust and the radiation and solar wind coming from the Sun. After the comet makes its closest approach to the Sun, the CRAF spacecraft will make a 50,000-km excursion down the comet's tail to explore the comet-solar wind interaction in more detail.

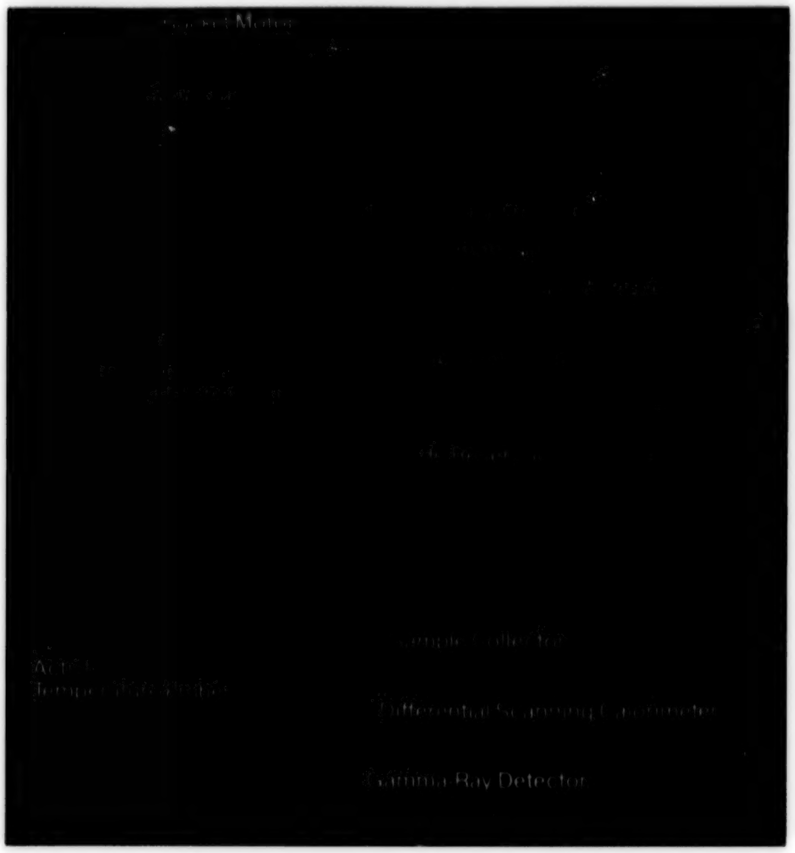


Figure 16-7. This cutaway diagram shows the various components of the CRAF penetrator, a harpoon-like probe that will bury itself in the comet's nucleus and transmit data to the spacecraft. Instruments that will measure the physical and chemical properties of the nucleus include accelerometers, gas analyzers, temperature probes, a gamma-ray detector, and a calorimeter to analyze samples collected from the surrounding nucleus.

The next Mariner Mark II mission after the CRAF Mission is called Cassini, and it combines two of the original Core Program missions recommended by the SSEC: the Saturn Orbiter and the Titan Probe. Cassini will explore the whole Saturnian system, which contains a host of volatile-rich objects—Saturn, Titan, many small moons, and the rings themselves (fig. 16-8). In addition to providing possibly pristine samples of the outer solar system, these objects may also contain records of the processes that have modified them. Like Galileo, the Cassini Mission will consist of two connected spacecraft, a Saturn Orbiter spacecraft built by

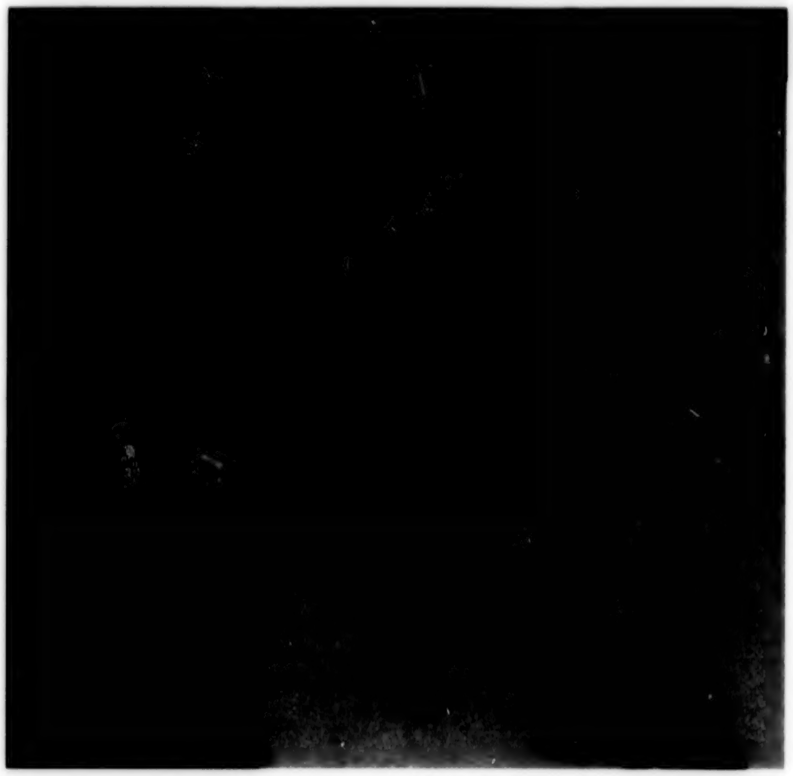


Figure 16-8. Saturn and two of its moons, Tethys and Dione.

NASA and a detachable Titan Probe, which might be supplied by the European Space Agency. This mission will enable us to begin to make comparisons between Saturn and the data we receive about Jupiter from the Galileo Mission.

Before going into orbit around Saturn, the Orbiter will deliver the probe to Titan. The short-lived probe, powered by batteries, will plunge into Titan's thick, organic-rich atmosphere, make chemical measurements as it descends, and broadcast the data back to the Orbiter until it reaches the hidden surface of Titan. Subsequently, the Orbiter will make close flybys of Titan on every orbit to carry out intensive studies of this most unusual moon, and a radar instrument will penetrate the opaque orange atmosphere of Titan to map the surface beneath it.

One of the most intriguing aspects of Titan is the possibility that its surface may be covered with oceans or lakes of liquid hydrocarbons produced by photochemical processes in Titan's atmosphere. These processes may be similar to those that occurred on Earth in primordial times, also producing organic chemicals that, in the more benign environment of Earth, led to the origin of life.

One of the benefits of the Cassini trajectory from Earth to Saturn is that it takes the spacecraft through the asteroid belt. By that time, Galileo and CRAF will have made close flybys of asteroids, and a different compositional type will be chosen for a Cassini flyby. Thus, in the course of these three great explorations, several asteroids, each with different surface compositions, will be studied independently.

The continued exploration of the outer solar system, where the United States has established preeminence, depends on the Mariner Mark II program and on the early approval of CRAF and Cassini to get it started. This major step had already been postponed for several years beyond the date recommended by the SSEC, and the CRAF Mission has become increasingly well-studied and well-defined. The inclusion of the CRAF and Cassini Missions in the FY 1992 budget is a welcome sign; this project is long overdue, and our continued exploration of the solar system depends critically on it.

Further delays in making such commitments will cause significant losses to our plans and our prospects for the future. We will lose the benefits, in cost, in planning, in science, and in morale, that come from being able to follow the Galileo launch with a strong and well-planned program of further exploration. We will lose the opportunity for significant cooperation with potential international partners, such as the European Space Agency, who have their own schedules and plans.

Finally, if we delay further, the movements of the planets themselves will work against us. In just a few years, Jupiter will move out of the alignment that now makes it possible to use it as a "slingshot" to reach Saturn in greatly shortened times. Without the ability to use Jupiter in this way, travel times to the outer solar system will be substantially longer. Further delay will, for all these reasons, create a major gap in the U.S. planetary program, a gap similar to the interval between 1978 and 1989, when no planetary missions had been launched.

The leadership that the United States has established in exploring the worlds beyond Mars deserves to continue, and it can. The outer solar system is waiting with its unknown worlds, its unsuspected discoveries, and its new knowledge. We are ready to explore it, and we know how to go about it. We can maintain visible national leadership here, and at the same time, we can push the frontiers of scientific discovery even farther out into the universe.

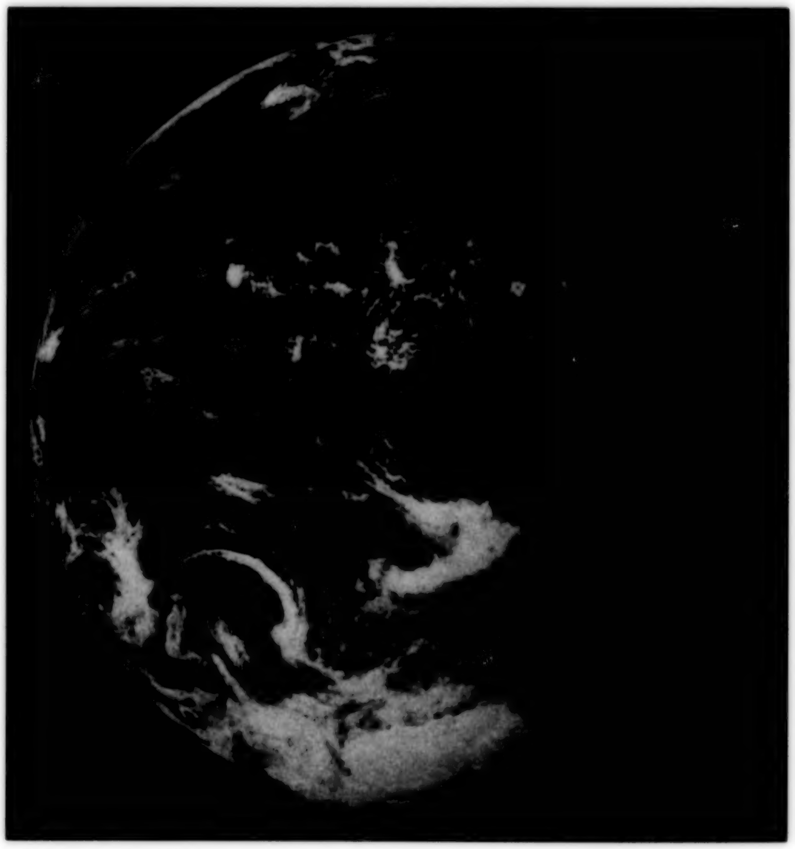
Comparing Planets: The Inner Solar System

Unlike the outer solar system, the worlds of the inner solar system, Mercury, Venus, Earth, the Moon, and Mars, have had much of the record of early solar system history erased from them. The inner planets have been dramatically changed from what they were originally, largely by their own internal forces. Radioactive elements trapped within them generated heat, and this heat produced geological activity, volcanism, earthquakes, mountain-building, and the development and migration of huge crustal plates. Original atmospheres were lost, perhaps stripped by the force of huge solar flares, sometimes replaced by volatiles released from inside, and then, on worlds of low gravity, slowly lost again to space. The atmospheres that did remain were continually modified by the Sun's energy, by internal volcanism, and, on Earth, by the development of life.

We study the inner solar system for many reasons, but for one reason in particular: to better understand our own planet by determining the present nature of similar worlds and by determining the general processes that govern all planetary development (fig. 16-9). The worlds of the inner solar system share many similar characteristics, but they also display a wide variety of individual features. The terrestrial, or Earth-like, bodies share common origins and compositions, and they have been modified by the same processes—meteorite bombardment, core formation, volcanism, mountain-building, and the formation of atmospheres. But these forces have acted differently on different bodies, and our search through the inner solar system is an attempt to relate what we see to the different processes that have produced it.

To understand these similarities and differences, we need to obtain a wide range of data. To fully understand the worlds of the inner solar system requires many extensive and difficult studies: global characterization of surfaces and atmospheres, use of geophysical stations for studying internal structure, and ultimately, the return of samples to Earth to determine their detailed character and to unravel the complex historical records that are preserved in them.

Figure 16-9. A view of Earth taken from an Apollo spacecraft.



A new impetus for studying the inner solar system comes from the recent announcement of a national goal to expand human presence beyond Earth orbit and into the solar system. The first and most likely locales for human exploration are the Moon and Mars. From now on, data gathered from these two worlds will serve a two-fold related purpose: to understand the Earth through comparative planetary studies, and to develop the technical and scientific foundation for future human exploration.

During the past two decades, significant strides have been made in the exploration of the inner solar system. The currently scheduled missions, Magellan and Mars Observer, will complete the global characterization of Venus and Mars. The next logical step is the intensive study of Mars by a sample return mission, a mission that will also significantly support the planning of future human initiatives to the Red Planet.

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The Planetary Observer program is the backbone of our strategy for the global characterization of the inner planets. The Mars Observer, planned for launch in 1992, will study the surface and atmosphere of Mars (fig. 16-10). The Lunar Observer, the next in the series, is under development to study the geoscience of the Moon.

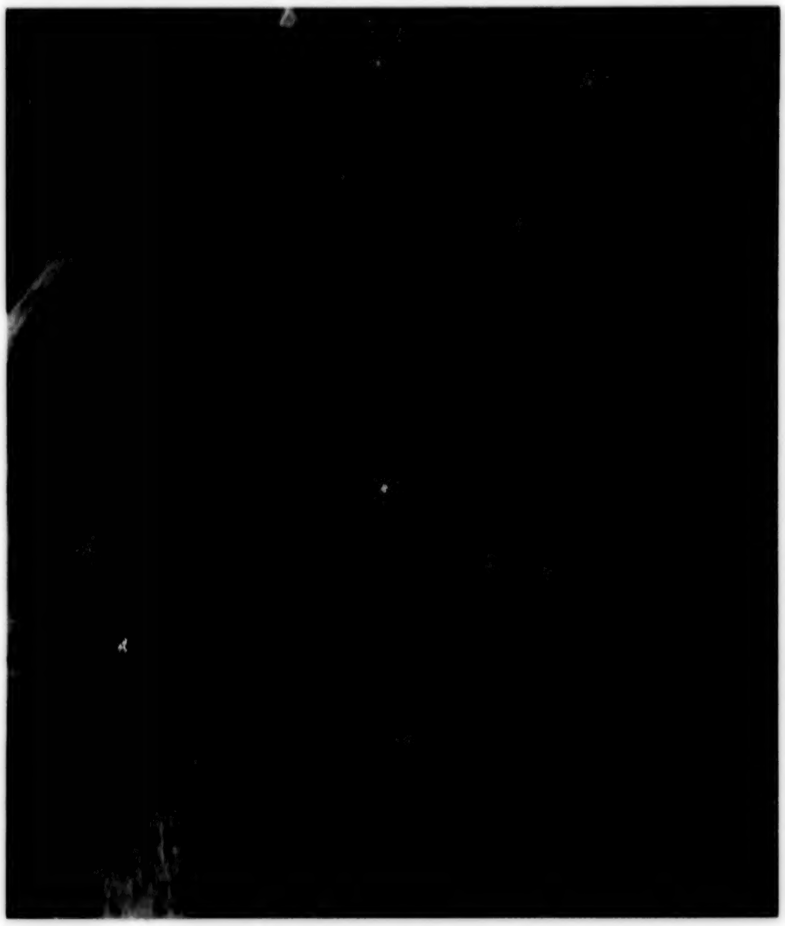

Figure 16-10. The panoramic view and complex geology of Mars is presented in a mosaic of computer-enhanced Viking pictures. The huge canyon of the Vallis Marineris stretches horizontally across the center of the picture, and three of Mars' huge volcanoes appear as brown spots at the left.

Beyond the Observers, a Mars Rover/Sample Return Mission is being studied—a mission that would land on the surface of Mars, traverse significant distances over that surface, and return a handful of Mars to Earth laboratories for analysis. Such samples would be chosen with the interests of both planetary scientists and exobiologists in mind, and they would provide a new level of understanding about Mars including its chemistry, mineral composition, physical properties, history, organic compounds, and the existence of present or past life.

The Mars Rover/Sample Return is a complicated mission. Much of the technology that would enable or enhance this mission does not yet exist; however, the recent funding for Project Pathfinder, both in FY 1988 and FY 1989, is encouraging. Many of Pathfinder's technology studies would support this mission. Because of the complex requirements, and also because the C.I.S. has stated its interest in such a mission, a Mars Rover/Sample Return is a potential candidate for a cooperative venture with the C.I.S. A NASA delegation traveled to Moscow in 1988 to discuss the process whereby cooperative exploration missions with the U.S.S.R. might occur. Cooperative discussions continue today with the C.I.S. We are also working in close cooperation with NASA's Office of Exploration to determine the requirements and plans for this mission relative to their plans for human exploration.

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Next Steps for the Future

The first priority for planetary exploration is to preserve the current missions and other elements of the ongoing program, and to ensure that scheduled missions are launched as soon as possible. In the meantime, we will begin to develop the payloads that will allow us to make maximum use of the Space Station: the Cosmic Dust Collection Facility, and the Astrometric Telescope Facility to search for other solar systems. Furthermore, we plan to augment the research and analysis programs that are so fundamental to the success and value of our flight projects.

Planetary exploration, in the 1990s, is facing a contradictory "best of times and worst of times." Many challenges and stresses in our program need to be overcome. But tremendous potential can be seen in future missions. We are on the threshold of major steps to begin exploring the outer solar system, and thus to discover our origins. We are poised to make major contributions to the national goal of major activities in space. We are planning and considering cooperative ventures with many other spacefaring nations. The possibilities are endless—the other worlds are waiting for us—we have only to choose to explore them.

Additional Reading

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